EVAPORITE-BEARING SEQUENCES IN THE ZECHSTEIN AND SALINA BASINS, WITH A DISCUSSION ON THE ORIGIN OF THEIR CYCLIC FEATURES

Peter Szatmari

University of Edinburgh, Grant Institute of Geology

TABLE OF CONTENTS

S U M M A R Y

Factors controlling cyclic sedimentation are discussed in a parallel study of two evaporite-bearing sequence, the Zechstein of Germany and the Silurian Salina Group of the Appalachian Basin. The Zechstein sequence was deposited in a basin that had received the debris swept in from the Variscan orogenic zone. The deposition of the evaporite-bearing sequence took place during a period of tectonic calm, preceded and succeeded by mild late Variscan movements. The sequence is divided into four major cycles by shale horizons accompanied and basinwards partially replaced by dolomites and anhydrites. Halite is the dominant sediment, it contains beds of anhydrite and potash salts, less commonly of shale, forming with the halite sedimentary cycles of diverse magnitudes.

The Salina Group has been deposited in a basin that had previously received debris from the Taconic orogenic zone. The last orogenic movements had virtually ceased before the deposition of evaporites commenced. The evaporite-bearing sequence is divided into three major cycles by shale suites related to alluvial fans of debris swept in from the previous orogenic zone. The shale beds are accompanied by dolomite beds containing stromatolitic horizons. The salt contains shale and dolomite beds of diverse thicknesses, giving rise to cycles of varied magnitudes. With increasing distance from the orogenic zone, the thinner shale interbeds in the salt grade into anhydrite.

In contrast to the Zechstein sequence, in the Salina Group thicker anhydrite beds are rare and no potash zones have been found. The anhydrite deficiency is attributed by the author to bacterial reduction of the $CaSO_4$. The H_2S thus formed is in part retained in the sediments, in part it deposited FeS_2 or re-oxidized. The lack of potassium salts indicates a less inhibited communication with the open sea, as also witnessed by repeated incursions of marine fauna.

In both sequences, most sedimentary cycles are controlled by the periodic entrance of diluted waters into the basin. Rain water enters directly as well as in the form of terrestrial run-off from the adjacent mountains, introducing mud and foreign ions, diluting and changing the ion ratios of the brines. Sea water enters the basin continuously or periodically; the concentration increases caused by the concomitant inflow of dissolved salts are mitigated by the reflux of more concentrated brines. Abrupt dilution of the brines by sea water followed by slow evaporation produces cycles of progressive solubility in the sediments resembling experimental successions.

The periodic entrance of rain and sea water can be controlled by several factors. Increases in rainfall, particularly in the detritus source area, may reflect morphologically or astronomically induced climatic changes; the morphologic factors may in turn be controlled by tectonism, erosion and sediment accumulation. The ingress of sea water can be caused by intermittent subsidence in the bar area, or by a rise of sea level induced tectonically, glacio-eustatically, or simply by a change in wind direction. A few models involving parallel control of terrestrial and marine inflow are presented at the end.

Introduction

On the following pages, two cyclic evaporite-bearing sequences, distant from each other both in time and in space, (the Permian Zechstein of NW Europe and the Silurian Salina of NE North America), will be described and the processes responsible for their cyclic structure will be analysed. It is hoped that such a parallel presentation will, like the parallel biographies of Plutarch, stress the common nature of distant happenings and deepen the validity of the conclusions.

The entertaining qualities of Plutarch's Lives are not expected to be parallelled. Apart from the not negligible differences between the literary abilities of the two writers, it is regrettable but true that human strivings, failings and failures arouse more interest in us than even the best anhydrite-shale contact. But the image of the eternal movement, the Heracleitic παντα ρει, may emerge and impress us equally whether the moving objects be animate or inanimate.

Though one cannot enter the same river twice, it is impressive that rivers, coasts and plains, remarkably reminiscent of their predecessors, repeatedly recur. While our interpretation of the causes of this recurrence may always be far (although hopefully not always equally far) from the real causes, it is the coherent reflection of natural processes in the mind, and not the physical processes themselves, which form the proper subject of science. This reflection changes as our attention focuses on different aspects of the physical process: like our eyes, our mind cannot focus simultaneously on two different objects with equal acuity. Important phenomena, seen clearly before, fade from our view, and relatively insignificant facts are accorded greater importance as they approach our shifting focus. But the fading images do leave their traces behind, so while these focal shifts create an infinite diversity of opinions and interpretations, changing from time to time and from person to person, and giving rise to debates between opinions equally well founded on both sides, it can be hoped that with these shifts the total knowledge deepens.

The more fleeting aspects of interpretation render it desirable to present separately the stratigraphic evidence and its sedimentological interpretation. While what we may be justified today in viewing as a deep-sea deposit, may tomorrow be even more justly considered a mountain-top deposit on the basis of newly found evidence, an anhydrite bed occurring at a certain position in the sequence and having certain structural characteristics will continue to exhibit most of these features for many aeons to come. As this separation of actuality from truth, anatomy from life, may somewhat damage the enjoyability of this writing, the reader may be well advised to gain his inspiration from the pictures rather than from the text of the descriptive parts, and concentrate his attention on the more analytical discussions.

Though one tends to feel one's opinions one's own, this is seldom the case: had I been born a few hundred years ago in India and exposed to the same factual evidence, I would probably have considered it an adorable trick of Maya, the goddess of illusion, rather than a product of repetitive sedimentation. Apart from the general empiricist-evolutionist intellectual milieu of our age, the following authors have greatly shaped my ideas:

F.H. Stewart, emphasising the value of microscopic observations, the complex genesis of mineralogical cycles, and the signficance of replacements in evaporites.

G. Richter-Bernburg, stressing the regularity and far-reaching correlation of evaporite-bearing cycles.

P. McL.D. Duff, focusing on the ways in which sediment accumulation alone induces self-controlled cyclicity.

A.V. Carozzi, D.J. Shearman and D.J.J. Kinsman, emphasising the significance of coastal and supratidal processes in carbonate and evaporite formation.

L.N. Strakhov, presenting arid sedimentation as a coherent system of interconnected processes leading from detritic to evaporite sedimentation.

Acknowledgments

For their help the writer would like to express his gratitude to Professor F.H. Stewart, Fellow of the Royal Society, Regius Professor of Geology at the Grant Institute of Geology, University of Edinburgh, for supervising this work and greatly developing my critical abilities;

Dr. P. McL. D. Duff, Grant Institute of Geology, for supervising, guiding and counselling my studies;

Professor G. Richter-Bernburg, President, Federal Geological Institute, Germany, for spending much of his free time introducing me into his research on Zechstein evaporite cyclicity and for arranging my field trips in Germany;

Mr. C. H. Jacoby, Manager, International Salt Company, for his kind help and support in both the United States and abroad;

Dr. Roth and Dr. Bauer of the Kali u.Salz A.G., Dr. E. Hofrichter, Dr. W. Kosmahl and Dr. P. Simon of the Federal Geological Institute of Germany for their guidance in Zechstein salt mines and for making their Ph.D. theses available;

Mr. Z. S. Szyprowski of International Salt Company, for his guidance in the Cleveland salt mine and for many stimulating discussions

Mr. D. J. Shearman, Imperial College, London, for the edifying discussion of his research on recent evaporite formation;

Dr. T. Scoffin, Grant Institute of Geology, for making me acquainted with his research on modern carbonate sedimentation in the Bahama Islands;

Mr. M. Saunders, Grant Institute of Geology, for making essential chemical analyses;

Mr. J. R. Bollar, International Salt Company, for his technical assistance;

THE PERMIAN ZECHSTEIN GROUP OF GERMANY

Tectonic-paleogeographic background

The Zechstein Sea (Fig. 1,17) was an intracontinental sea in Northern Europe, resembling in some of its aspects the present North and Baltic Seas. To the south, it was bordered by the still active Variscan Mountains, stretching from Ireland through Cornwall to the present Rheno-Ardennian and Vincelician-Bohemian Mountains. To the west, it was terminated by the eroded remnants of the Caledonian Mountains (and perhaps the Canadian Shield beyond them); to the east, it intruded deeply into the flat Fennoscandian-Russian Shield. Along its axis, it was about 1800 km long, stretching from Yorkshire to the Baltic countries. Occupying the North Sea basin, it was presumably open to a Boreal Sea in the north. Whether it was connected with the Tethys through the Carpathian geosyncline is unknown; generally it is thought that no such connection existed.

The sea entered from the north at the start of the Late Permian, invading a rapidly subsiding basin. During earlier Permian and Carboniferous times, the subsidence was being compensated by clastics, swept in by rivers coming from the Variscan Mountains in the south and west, and deposited first as coal-bearing sequences, later as mainly fluviatile red beds. The drainage system of these early Permian rivers strikingly resembled that of the present rivers of the German-Polish lowlands; a southward ingression of the North and Baltic Seas into the adjacent plains would produce a geographic pattern very much like that of the Zechstein.

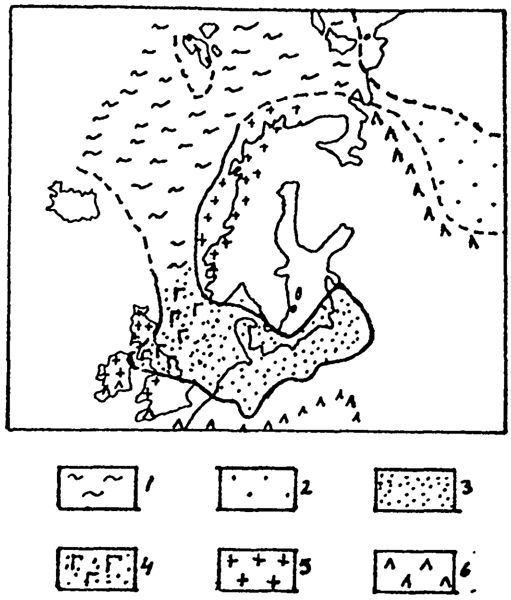


Fig. 1. General palaeogeography of Zechstein (Upper Permian) time. (From Stille, 1932, and others). 1. Open boreal sea; 2. Russian Zechstein sea; 3.German Zechstein sea; 4. Bar area: Caledonian structures under the sea; 5. Eroded Caledonian mountains; 6. Variscan mountains.

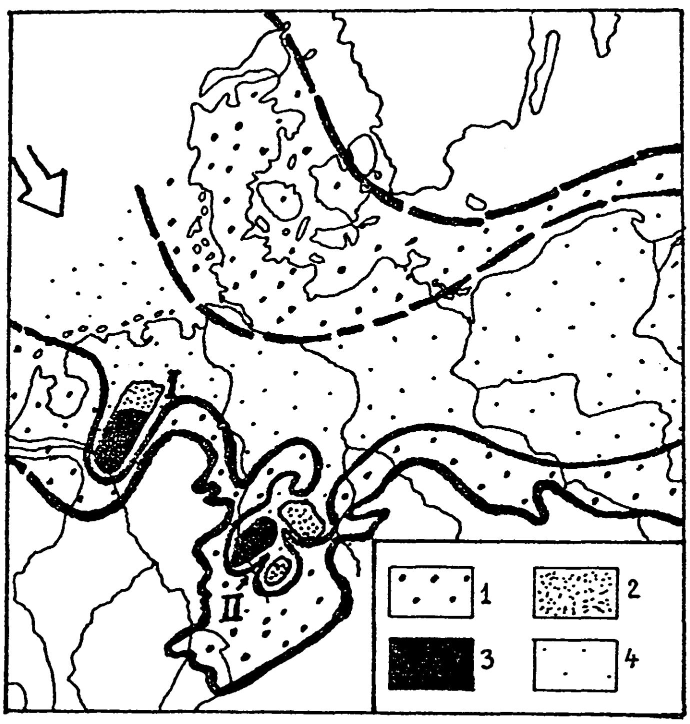


Fig. 2. Palaeogeography of the Werra time (after Richter-Bernburg, 1957; Strakhov, 1970). 1. Massive carbonates and anhydrites along the basin margin; 2. Halite; 3.Potash salts; 4.Laminated dolomite and anhydrite in the basin centre; I - Lower Rhine Basin; II - Werra-Fulda Basin.

The entrance of the sea marked a temporary slowdown in the uplift of the Variscan Mountains in the south; at the end of the Permian, when the tectonic activity resumed, the sea again withdrew, giving place to renewed deposition of detritic sediments in a continental environment. (A similar period of tectonic tranquility was responsible for the deposition of the upper Silurian - lower Devonian carbonate-evaporite suite of the Appalachian basin (see Part II) between under- and overlying molasse sequences).

The rift that was to widen into what is now the Atlantic Ocean (Bird and Dewey, 1970) was scarcely open during the late Permian, and communication with the Boreal Sea was inhibited by remnants of the Caledonian Mountains submerged between Scotland and Norway, leading to the formation of a basin with restricted circulation, favorable for avaporite deposition. Surrounding of this marine basin by continents at arid paleolatitudes (20-30°N) provided appropriate climatic conditions , while the relatively rapid subsidence concomitant with the rise of the Variscan Mountains (and possibly related to the opening of the Atlantic Ocean) still continued. In such conditions, the land-locked sea was filled principally with evaporites, deposited from presaturated sea water. These brines entered the basin overflowing the "bar" of the submerged Caledonids in the North Sea region, and were periodically diluted by occasionally less inhibited sea water inflow as well as by rain falling directly into the sea or brought in by rivers from the more humid Variscan highlands.

The oscillation of marine inflow, and recurrent periods of higher precipitation, produced a sequence of repeated sedimentary cycles. In view of the scarcity of fossils, Lotze (1938) indicated the feasibility of building a lithostratigraphy based on these cycles. This work was accomplished by Richter-Bernburg (1941, 1953), who differentiated four major cycles and correlated them throughout the German part of the basin. Stewart (1954, 1963) and Lotze (1958) extended this correlation to include the three major cycles of England.

Each of these main cycles consists of two elements. A thick progressive sequence grades from shales through carbonates and anhydrite into halite and eventually into potash salts; a thin, sometimes missing, recessive sequence leads back to shales through the same phases in reverse order. Minor oscillations complicate the stratigraphy.

During the last twenty years, attempts have been made to use the smallest, frequently annual, cycles within individual anhydrite and halite suites for fine stratigraphic work (Richter-Bernburg, 1957, 1960; Hofrichter, 1960). One of the major contributions of this method has been to show that some of the lithostratigraphic boundaries are not strictly time-parallel, but separate partially time-equivalent, sometimes lenticular bodies. The boundaries between the cycles, on the other hand, appear to be isochronous.

In the following discussion, we shall follow the cyclostratigraphy of the Zechstein, presented below after Richter-bernburg (1953) and used both by science and industry. While reviewing the entire Zechstein, we shall concentrate on units providing most information on cyclic sedimentation.

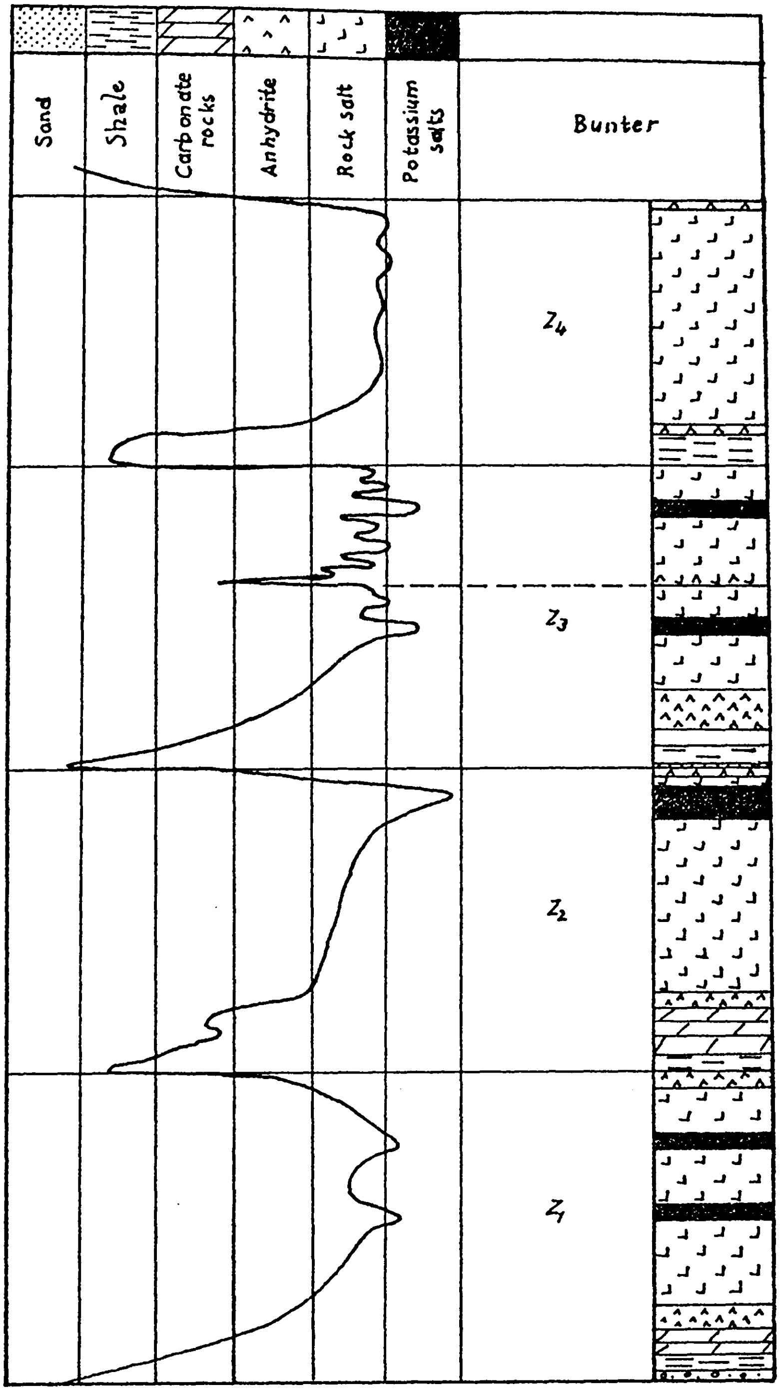


Fig. 3. The cyclic subdivision of the German Zechstein (from Richter-Bernburg,1957). Z_1 Werra; Z_2 Stassfurt; Z_3 Leine; Z_4 Aller Cycle.

Stratigraphy of the German Zechstein (Fig. 3)

After Rechter-Bernburg, 1953, abridged

ZECHSTEIN 4		Salt with clay laminae
Aller cycle		Salt with clay fragments
		Pink salt
		White salt
		Pegmatite anhydrite
		Red shale
ZECHSTEIN 3	Riedel Group	Salt with clay laminae
Leine Cycle		Riedel potash deposit
		Swath salt
		Salt with anhydrite interbeds
	Ronnenberg Group	Variegated salt
		Banded salt
		Layered salt
		Ronnenberg potash deposit
		Lineated salt
		Anhydrite crust
		Black shale lamina
		Main anhydrite
		Magnesite
		Gray shale
ZECHSTEIN 2		Cover anhydrite
Stassfurt Cycle		Cover salt
		Stassfurt potash deposit
		Stassfurt halite
		Basal anhydrite

ZECHSTEIN 2	Main dolomite-
Stassfurt Cycle	Bituminous shale
	Brownish red shale
ZECHSTEIN 1	Upper anhydrite
Werra Cycle	Upper salt
	Hessen potash deposit
	Middle salt
	Thüringen potash deposit
	Lower salt
	Lower anhydrite
	Carbonates
	Copper shale
	Conglomerate

Z1 - WERRA CYCLE

The first cycle of the marine Zechstein covers and transgresses beyond the limits of the lower Permian fluvial red sandstones, overlying rocks ranging from Silurian to lower Permian. To the southwest (Werra-Fulda and Lower Rhine basins, Fig. 2-3), it forms rather narrow bays, intruding into the Variscan Mountains. The fullest profiles of the Werra Cycle are found in these subbasins:

Upper rock salt	-100 m
Hessen potash horizon	3 m
Middle rock salt	60 m
Thüringen potash horizon	3 m
Lower rock salt	100 m
Anhydrite	10 m
Limestone	8 m
Copper shale	.5 m
Conglomerate	5 m
Lower Permian red beds	

Conglomerate

In a narrow belt (200-300 m) along the basin margin, the base of the Zechstein is marked by a conglomerate zone. It is normally only a few metres thick and consists chiefly of gravel-size clasts, though larger pebbles and boulders occasionally also appear.

Copper Shale

The overlying shale is mostly black, and contains sulphides of copper, lead and zinc in economically significant amounts. The copper content increases to about 1% along the basin margin, but drops again at the very margin where the black copper shale grades into uneconomic red shales.

The black shale contains about 10% of organic carbon at its base, which decreases upwards as the carbonate content increases. Because of its economic significance, the thin (-0.5 m) Copper Shale has been subdivided into several horizons (Fig. 4). The fact that the lower horizons frequently wedge out where the Copper Shale overlies eolian sand dunes along the basin margin, indicates that the shale was formed in gradually ingressing shallow sea.

The origin of the base metal concentrations has been the subject of many scientific debates. Now it is most commonly believed that uneconomic syngenetic sulphides were mobilized by late diagenetic hot ("telethermal") brines and redeposited in the most favourable geochemical environment.

Carbonates

As in following Zechstein cycles, the basal detritic sediments soon gave place to carbonate deposition. Thick reefs mark the outlines of the basin, overlying both pre-Permian and lower Permian rocks. Occasionally, the basement rocks form abraded cliffs, with huge detritic blocks filling intracliff hollows; the overlying reefs grow equally on the abraded

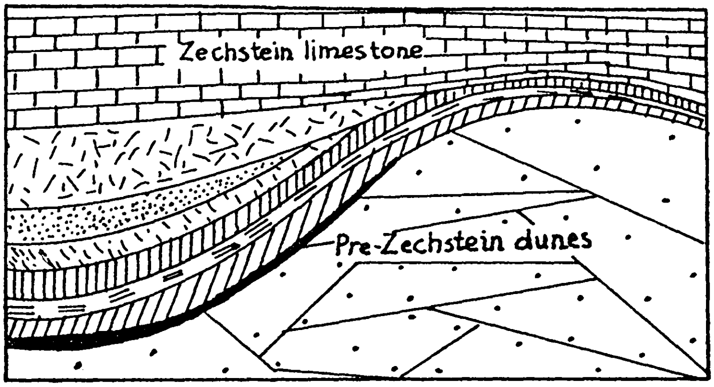


Fig. 4 . Wedging out of divers copper shale horizons between the Zechstein limestone and the pre-Zechstein dunes. Theickness of profile: 0.6 m. (After Kautsch, 1942).

basement and over the abrasional boulders (Richter-Bernburg, oral comm., 1970). Non-reef carbonates of the marginal regions are massive, light-colored and bioclastic, locally also oolitic; they are rich in brachiopods, pelecypods, and bryozoans. Basinwards they pass into thin-bedded, bituminous marls, with the disappearance of shell debris. In the central regions, these marls are only about 4 m thick.

The thick marginal carbonates grade basinwards, the thin dolomites of the basin interior grade upwards into anhydrites (Fig. 5). Close to the margin, a transitional anhydritic dolomite occurs, to which leaching of sulphates gives a froth-like structure in surface exposures.

Anhydrites

The anhydrite is over 100 m thick along the margin, thins to 5-10 m in a discontinuous belt inwards from the marginal belt, reaches again thicknesses of 400 m in an inner "barrier" belt, and is again thinner (40-100 m) over the interior where it contains extensive halite lenses.

The structure of the anhydrite is controlled by its position. Thick, massive, light-colored beds occur in the inshore belt, while laminated bituminous varieties, locally interlaminated with dolomite, cover the similarly bituminous laminated carbonates over the interior of the Zechstein basin.

Especially the marginal anhydrite frequently displays lenticular "flaser" structure: anhydrite nodules several

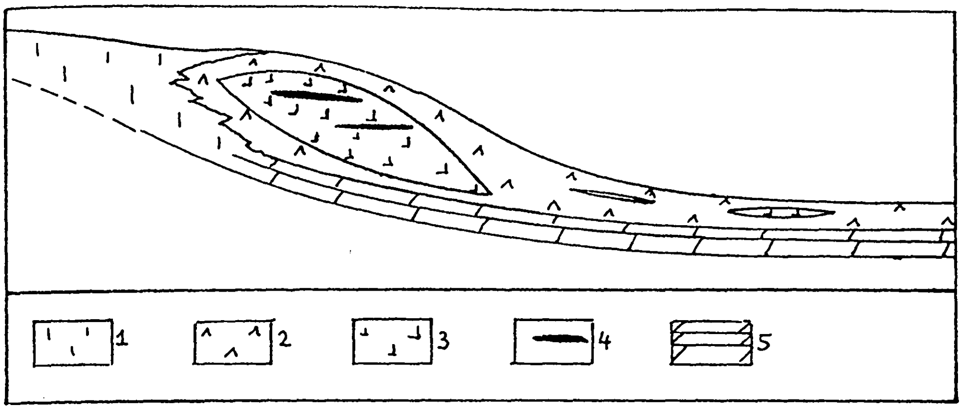


Fig. 5. Facies pattern of the Werra Cycle. 1. Marginal carbonates; 2. Anhydrite; 3. Halite in marginal depressions; 4. Potash salt deposits; 5. Laminated carbonates. (Modified after Richter-Bernburg, 1957).

cm large are surrounded by wisps or flasers of bituminous carbonates (chiefly dolomite and magnesite). Truncated cycles, in which flaser anhydrite grades upwards into laminated anhydrite are common; their thickness is 5-10 m. The varved anhydrite of each cycle is cut off by an unconformity surface, to be overlain by the flaser anhydrite of the subsequent cycle (Richter-Bernburg, oral comm., 1970).

Halite

The distribution of halite in the Werra Cycle differs substantially from its distribution pattern in the overlying Zechstein cycles. While in the latter it occupies the interior, here the interior is essentially salt-free and the major salt deposits occur near the margin, between the marginal and the "barrier" anhydrite belts mentioned before. The three major salt deposits occur in the Lower Rhine, Werra, and Fulda subbasins (Fig. 2).

In the Werra basin, where it is well exposed in the Wintershalle Mine of the Kali u. Salz A.-G. at Herringen, the salt is subdivided into three sections, separated by two potash horizons.(The standard sequence is presented at the beginning of the discussion of the Werra Cycle).

Lower rock salt

The halite overlying the basal anhydrite is pale; salt beds of more or less uniform thickness are separated by thin laminae of argillaceous anhydrite (Fig. 7). An internal la-

Figure 6

Internal lamination in halite. The fine oscillation of sylvitic-haematitic red and pure pale halite laminae may reflect annual cycles; the argillaceous-anhydritic gray beds may be 11-yearly deposits.

Lower Werra halite, Herringen, Germany

Figure 7

Pale gray halite beds, separated by argillaceous anhydrite laminae (dark gray). Some internal lamination (annual?) is vaguely perceptible within the halite beds

Numbers on both pictures indicate depth in meters below the base of the Herringen potash deposit

Lower Werra halite, Herringen, Germany

Fig. 6

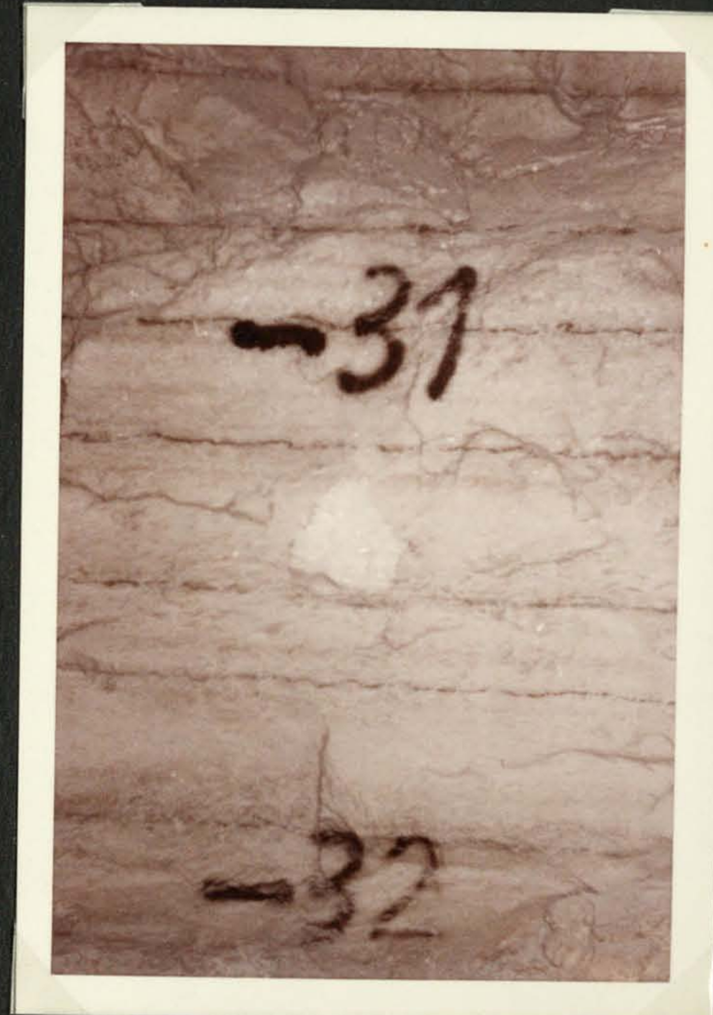


Fig. 7

mination is vaguely perceptible within the halite beds.

In the upper quarter of the Lower rock salt, this internal lamination becomes very distinct (Fig. 6). Between the thin (<1 cm) anhydrite laminae, the halite beds are 10-25 cm thick. Within each halite bed, thicker pale and thinner red laminae alternate; the red laminae contain several per cent KCl and dispersed haematite. The higher potassium and bromine (280 ppm) content of this horizon marks its transitional nature into the overlying Thüringen potash deposit.

The number of pale and red laminae is remarkably constant; in each halite bed it totals about 11, ranging from 9 to 12. The causes of this double cyclicity (argillaceous anhydrite alternating with halite and within the halite, potassiferous laminae alternating with pure ones) is strongly debated; like many newer observations, it seems to be in contrast with the previous assumption that each halite bed was deposited within a single year. We shall return to this problem in a later chapter.

<u>Middle rock salt</u>

Above the Thüringen potash deposit, the bromine content stays at a high level (240-280). The anhydrite lamination is distinct, and even the internal lamination is often recognizable within the halite beds. Though the lamination is normally parallel and even, there are two zones of disturbed lamination. In the thicker one, between 42 and 14 m below the base of the Hessen potash deposit, the lamination is chaotically disturbed. In the upper one, between 7 and 6 m below the Hessen deposit, the varves between the halite beds consist of langbeinitic-kieseritic halite, and form a broken horizon both under and overlain by quiet lamination. Presumably, the zones of disturbance represent underwater mudflows or short-lasting leaching in the surface layer of the sediment.

Upper rock salt

Above the Hessen potash deposit, the composition of the varves slightly changes. The upper 1-2 cm of each halite bed contains sylvite and kieserite; these minerals continue in higher concentration into the argillaceous-anhydrite varves that separate the halite beds.

Z2 - STASSFURT CYCLE

Red Shale

The second major cycle of the Zechstein, the Stassfurt Cycle (Fig. 8) commences with brownish red saliferous shales grading into dark fetid shales toward the basin interior. Although this shale bed is generally very thin, it may represent a long period during which no evaporite was deposited (or no newly deposited evaporite was preserved) and an unknown amount of previously deposited evaporites was dissolved.

Main Dolomite

The distribution pattern of the economically very impotant, hydrocarbon-bearing carbonates of the Stassfurt Cycle resembles the distribution of the basal Werra carbonates. Along the margin, the algal reefs are up to 60 m thick. Toward the basin interior, much of the dolomite becomes replaced by anhydrite in a still relatively marginal zone. The lateral contact of dolomite and anhydrite is marked by a belt in which anhydrite concretions are very common in the carbonate rock; their early diagenetic leaching results in a cellular-cavernous, frequently brecciated carbonate structure.

In the basin interior, the dolomite is very thin (4-6 m), finely laminated, and often fetid; shale microlaminae and limestone interbeds are common. With increasing distance from the basin margin, the magnesium content gradually decreases and the anhydrite concretions disappear.

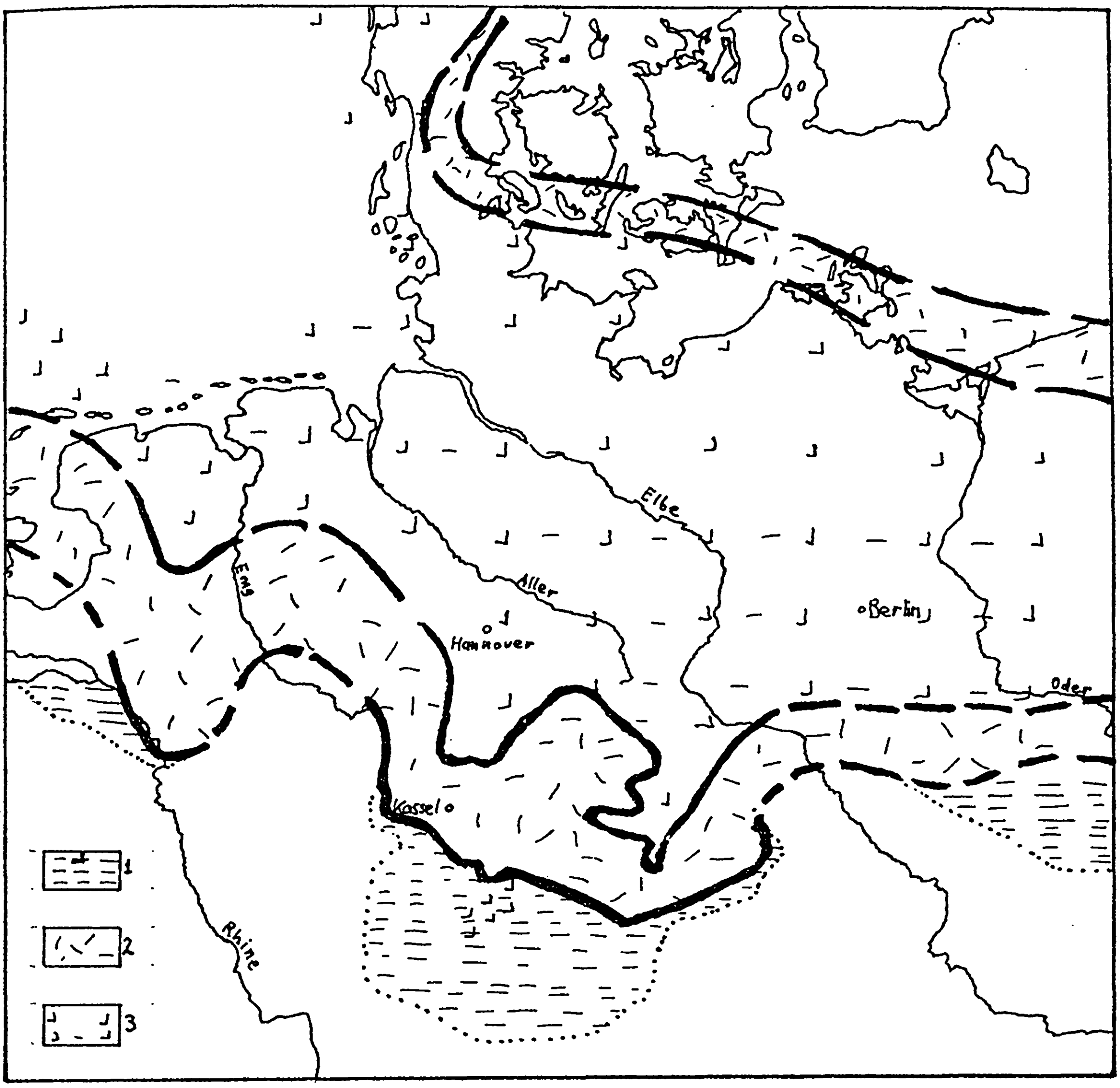


Fig. 8 . Palaeogeography of the Stassfurt Cycle. (after Richter-Bernburg, 1957). 1. Red pelites, locally with halite lenses; 2. Thick carbonates and anhydrite along the basin margin; 3. Halite, more than 200 m thick, with potash salts; partly underlain by fetid shales.

Basal anhydrite

Both upwards and basinwards the carbonates are replaced by anhydrite. A belt of anhydrite, up to 100 m thick, borders the thick marginal dolomite belt toward the basin. In the interior, the thickness of the anhydrite decreases to 0.8 m, as the rest of the anhydrite suite becomes laterally replaced by halite. This thin anhydrite, underlying the halite, resembles in its finely laminated structure and high bitumen content the underlying basal dolomite and occupies the same area, an area larger than the one covered by the similar lithofacies in the Werra Cycle.

In the interior, flat anhydrite lenses appear at the top of the dolomite, and these, gradually merging into congruous laminae, provide the transition into the basal anhydrite. In the marginal areas, no such gradation has been observed.

In the anhydrite, four major sedimentary structures have been distinguished. In the "lineated" or varved anhydrite, thin (0.2 mm) laminae of shale or dolomite alternate with thicker (2 mm) laminae of anhydrite. In the banded anhydrite, the thickness of the anhydrite laminae increases to 2-3 cm, and the non-anhydritic laminae between them become somewhat irregular. In the pearl anhydrite, small (1-10 mm) irregular lenses or concretions of anhydrite are arranged into indistinct laminae and are separated by a carbonate network. Finally, in the flaser anhydrite, the irregular lenses are several cm in diameter.

The laminae of the 1-m-thick anhydrite bed of the basin interior are well correlated (Richter-Bernburg, 1957, 1960). Even the presence of minor irregularities, the occurrence of one or two vague, indistinct carbonate laminae between two distinct ones, the straight, wavy, or finely zigzagged surface of the carbonate films, the occurrence of fragmented carbonate lamellae within an anhydrite lamina, can all be correlated, with an almost incredible accuracy, over a distance of 20 km.

If only the thickness of the anhydrite laminae is considered, the laminae can be correlated over distances of up to 200 km. For this purpose, the thickness of each varve is presented diagrammatically, and the diagrams thus obtained are readily correlated. The correlation of these curves is so remarkable, that even the position of a 10-cm core in the thin sequence can be unequivocally determined. A few laminae of flaser anhydrite occur within the lineated anhydrite throughout the basin; the ordinal number of the anhydrite laminae they overlie remains the same over a distance of 120 km.

The correlation of varves (about 800 altogether in the non-marginal anhydrite unit) has revealed important facies transitions. Toward the margin, first the thickness of the anhydrite laminae increases several times, without altering their proportions; then beds of flaser anhydrite intercalate; finally, in the marginal zone, the entire laminated anhydrite is replaced by flaser anhydrite. Towards the margin, the transition from laminated into flaser anhydrite occurs at progressively lower varves, indicating that conditions characteristic of the basin margin gradually advanced toward the interior of the basin.

In the interior itself, the thickness of the anhydrite laminae decreases to such an extent that here at the bottom of the anhydrite layer, in the vicinity of the underlying dolomite, they are often inseparable, limiting the application of the varve chronology. The countable varve sequence starts here 400 varves higher than in a near-margin zone 300 km to the south, indicating that it took at least 400 years for conditions favourable for varved anhydrite deposition to move 300 km.

Thus the anhydrite unit is underlain by dolomite containing anhydrite laminae and concretions; it starts with discontinuously laminated to flatly lenticular anhydrite in a flat dolomite network; grades upward into laminated anhydrite and ends with flaser anhydrite throughout the basin. The gradual transition of the discontinuously laminated anhydrite both upwards and towards the basin margin into first varved and then flaser anhydrite indicates a gradual shift of basin margin conditions towards the basin interior.

Stassfurt halite

In sharp contrast to the salt of the underlying Werra Cycle, the halite of the Stassfurt Cycle is not restricted to small marginal depressions, but covers the whole central area of the basin interior, extending beyond the limits of the underlying bituminous laminated anhydrite and dolomite. Occurring over an area of 350,000 km^2, in thicknesses up to 500 m, it is one of the largest salt deposits of the world. The halite of the interior extends into half-open marginal depressions to the south. Of these, the Weser Depression near Hannover resembles in its profile the main basin, while the sequence of the Sub-Hercynian basin is more aberrant and resembles the similarly marginal Thuringian basin.

Stassfurt halite of the Weser Depression

The halite sequence of the Weser Depression is well exposed in the Hansa Mine, near Ronnenberg. The major part of the halite is very light; the crystals range from 10 to 15 mm in size, with relatively few crystalloblasts up to 10 cm in diameter. The cyclic bedding (Fig. 9) consists of 15 to 20 cm beds of halite, alternating with beds and laminae of anhydritic halite, up to 7 cm thick. In the anhydritic halite beds, the anhydrite occurs in the form of dispersed flakes; the amount of kieserite is very small (about 1/20 of the anhydrite) and the proportion of argillaceous matter is still smaller.

In the top section of the halite, huge flat megacrystals,

Fig. 9

Pale gray halite interlaminated with dark argillaceous anhydrite and white kieserite. Numbers show the distance from the overlying Stassfurt potash deposit, measured in metres perpendicularly to the bedding.

Stassfurt halite; transition toward the overlying kieseritic zone, Hansa Mine, Hannover, Germany

Fig. 9

up to 15 cm large, occur frequently in the halite. The central part of these crystals is milky and displays a flat lamination which is not perfectly parallel to that of the enclosing halite. The bromine content is lower than in the enclosing rock. Locally, the cleavage network inside the megacrystals is cut off by a kieseritic lamina of the enclosing halite, indicating early diagenetic formation and partial dissolution.

Stassfurt halite in the Sub-Hercynian basin

The somewhat aberrant profile of the Stassfurt halite in the Sub-Hercynian basin, west of the Weser Depression, is well exposed in the Asse Mine and has been studied in detail by Simon (1970).

Anhydritic halite zone

The major part of the Stassfurt halite is similar to that of the Weser Depression. Beds of clear to milky salt, 5-10 cm thick, alternate with 0.2-2 cm anhydrite varves. The latter are light grey to black, according to the amount of argillaceous matter in them; locally finer, light and dark laminae can be distinguished within a single anhydrite lamina. Salt megacrystals, oval to oblong, are common; they are up to 5 cm large and frequently lie at the bottom of the halite beds forming discontinuous laminae. Megacrystals occurring in the middle of halite beds are frequently lighter than those occurring at the bottom of a bed, and sometimes display internal lamination.

While this major part of the Stassfurt halite resembled that of the Weser Depression and the basin interior, the uppermost 60 m below the Stassfurt potash zone shows important differences. Two main zones can be distinguished: a lower, polyhalitic zone is traceable throughout the Sub-Hercynian basin, while the upper, argillaceous zone has been noted only in the Asse Mine.

Polyhalitic halite zone

In the Asse Mine, a transitional layer, 0.5-2.5 m thick, leads from the "normal" anhydritic to the polyhalitic zone. This transitional layer consists of grey to white halite beds intercalating with cm-thick laminae of anhydrite and anhydritic polyhalite. Where the layer is the thickest, however, it is pale red in the upper 2 m and polyhalite laminae occur only at its bottom; upwards they are replaced by anhydrite laminae up to 5 cm thick.

The transitional layer is overlain by 22 m of relatively pure milky salt, in which indistinct laminae (1-3 cm) of polyhalite and slightly anhydritic halite occur at 10-30 cm intervals. The central zone of these laminae, about 0.5 cm thick, consists of almost pure, finely crystalline polyhalite. In the area where the argillaceous salt occurs (see later), this salt is thinner and more polyhalitic, consisting of three zones. Its bottom zone, 3m, is pale brown, and contains laminae of polyhalitic halite. The middle interval (1 m) is red, massive, contains dispersed polyhalite crystals and is overlain by an anhydrite lamina 3 cm thick. The top zone is 3-4 cm thick; it consists of coarsely crystalline salt with dispersed polyhalite.

The top portion of the polyhalitic zone contains the largest amounts of polyhalite. Here salt beds, 10-30 cm thick, alternate with yellow to red, finely crystalline polyhalite beds (1-20 cm) and with finely crystalline sulphate laminae in which either polyhalite or anhydrite is the dominant mineral. Halite is, however, still by far the predominant mineral even in this top portion of the polyhalitic zone: the average NaCl-content exceeds 91%.

Argilllaceous halite zone

In this local, 25-m argillaceous sequence of the Asse Mine, beds of grey to red halite are interlaminated with very thin, often black, slightly anhydritic shale laminae, occurring at 6-cm intervals. Sections in which the shale laminae are only a few mm thick and more loosely spaced alternate with intervals in which they are up to 2 cm thick and occur more frequently (Fig. 10).

Both the red color and the clay content of this zone resemble other shallow-water facies equivalents of the Stassfurt salt (Göttingen) and are thought to reflect the vicinity of a contemporaneous dry land. Their similarity to the lower Syracuse argillaceous halites of the Appalachian basin may indicate similar genesis. (see Part II).

Fig. 10

Light-colored halite, rhythmically alternating with slightly anhydritic, dark shale laminae

Near-shore argillaceous halite zone at the top of Stassfurt halite; Asse Mine, Braunschweig, Germany

Fig. 11

Gray halite interlaminated with dark greenish gray, slightly anhydritic shale and some white kieserite. The pink carnallite veins derive from the overlying potash deposit.

Near-shore argillaceous facies of kieseritic zone below Stassfurt potash deposit; Asse Mine, Braunschweig, Germany

Fig. 12

Pale pink carnallitic halite with dark greenish gray, slightly anhydritic shale laminae

Contact of Stassfurt potash deposit with the underlying argillaceous-kieseritic halite; Asse Mine, Braunschweig, Germany

10

11

12

Kieseritic halite zone

Throughout the German part of the Zechstein basin, a halite layer 2-5 m thick, containing very thin kieserite laminae at 8-cm intervals, forms the transition from the Stassfurt halite into the overlying Stassfurt potash deposit. In the Weser depression, the anhydritic halite grades into this kieseritic zone with a gradual shift in the anhydrite/kieserite ratio and increasingly light color of the salt. In those parts of the Sub-Hercynian basin where the argillaceous zone occurs (see above), the overlying kieseritic zone also contains shale laminae (Fig. 11).

Stassfurt potash deposit

The potash deposit developing from the kieseritic zone is the largest one in the world. It covers an area of about 100,000 km^2, or about the central third of the area occupied by the Stassfurt halite. In spite of the large area, alternating beds of halite and potash salts can be correlated, indicating overall control of the cyclic sedimentation.

This potassiferous zone is about 25 m thick. In the Weser depression (Hansa Mine) the lower part is relatively poor in potassium (3-5% KCl); it consists mainly of halite with sylvite, kieserite and boron minerals. The sedimentary cycles are distinct: 8-cm sylvitic halite beds alternate with kieserite laminae 0.5-1 cm thick. The proportion of the anhydrite is only about 1/20 of the kieserite. - The upper two-thirds of the deposit are richer in potassium, consisting of about 50% halite, 30% sylvite, and 15-20% kieserite. Anhydrite and boron minerals amount to less than 1%.

In the Asse Mine of the Sub-Hercycnian basin, the potash deposit consists chiefly of carnallite. Secondary veins of kieseritic carnallite, sometimes parallel with the lamination, mark the transition from the top, kieseritic-argillaceous zone of the Stassfurt halite into the overlying potash deposit. Near the base of the deposit here shale laminae are still common (Fig. 12). The low er part of the thick (50 m) deposit displays a cyclic interlamination of halite, kieserite and carnallite (Fig. 13), while in the upper portion pink carnallite dominates and a few thicker halite beds also occur.

The Stassfurt potash deposit has a distinct palaeogeographic pattern. The central regions of the basin, where varved bituminous anhydrite and dolomite underlie the Stassfurt salt, are characterized by carnallite with irregular lenses of kieseritic sylvinite (hartsalz). In the marginal belt, kieseritic sylvinite becomes predominant. At the southernmost edge of the basin, this kieseritic sylvinite is replaced by anhydritic sylvinite. This zoning, according to Lotze and Richter-Bernburg, favors the theory of primary depositional control of these minerological facies, even if we do not have to assume that the minerals deposited were the same minerals which are presently observable (e.g., probably epsomite was the primary antecedent of kieserite (Stewart, 1963)).

Fig. 13

Rhythmic alternation of dark greenish gray shale, pale greenish gray halite, white kieserite and red carnallite laminae. Stassfurt potash deposit, Asse Mine, Braunschweig, Germany

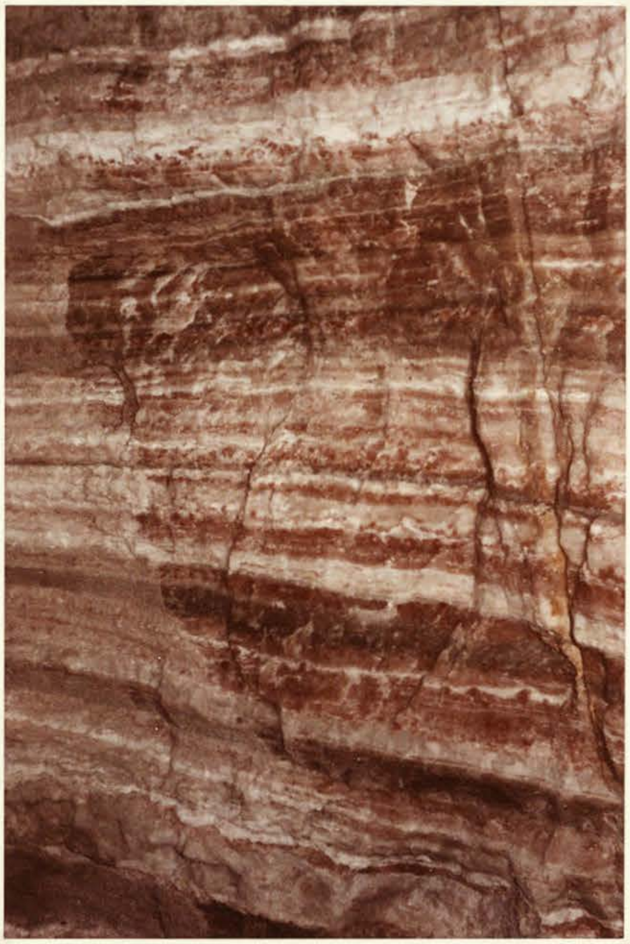


Fig. 13

Cover salt and cover anhydrite

In the Weser depression, the potash deposit is overlain by recessive evaporite deposits, marking the end of the Stassfurt Cycle. The lower portion of this recessive sequence is brown salt, less than 1 m thick, interlaminated with kieserite. The upper portion consists of anhydrite, about 1 m thick; its good lamination is due to the regular alternation of thicker (0.3-3 cm) pure and thinner (0.2-1 cm) argillaceous anhydrite bands; the laminae are flat to slightly wavy (Fig. 14).

Fig. 14

Alternating light gray pure anhydrite and yellowish gray argillaceous anhydrite - anhydritic shale laminae, providing the transition between the Cover Anhydrite at the top of the Stassfurt Cycle, and the overlying Gray Shale at the base of the Leine Cycle. Height of section: 2.4 m

Contact of Stassfurt and Leine Cycles, Hansa Mine, Hannover, Germany

Z3 - LEINE CYCLE

The basal sediments of the Leine Cycle, the Gray Shale, the Magnesite Unit, and the Main Anhydrite, have been studied in detail by Kosmahl (1969). In this discussion we shall follow his study, supplemented by our personal observations in the Weser Depression (Hansa Mine, near Hannover). To avoid nomenclatural confusion, we treat the non-argillaceous portion of the "Gray Shale" (Dolomitic Gray Shale) separately under the name of Magnesite Unit.

Gray Shale

The argillaceous base of the third major Zechstein Cycle, the Leine Cycle, overlies the uppermost sediments of the Stassfurt Cycle with a thin transitional sequence. (Fig. 14). In its thin basal section (0.3-1 m) the alternation of sandstone and shale replaces the underlying chemical cycles. The fine-grained sandstone laminae are gray, 1-30 mm thick; the argillite laminae are somewhat darker and thicker (1-80 mm). This basal section is overlain by dark gray to black sandy shale which locally contains "nests" of anhydrite, roughly vertical bodies a few cm high and a few mm thick. Previously these "nests" were thought to be anhydrite pseudomorphs after gypsum crystals, but recent research indicates that they fill discontinuous fissures which dissect the lamination at divers angles.

The thickness of the shale, and particularly of its most arenaceous basal section, increases toward the basin margin, especially toward the mountainous source areas in the south and east. In this direction, the gray shales become gradually replaced by red shales.

The muds which built up the Gray Shale Formation were carried into the basin by rivers coming from the southern highlands; eolian transport was found to have been subordinate. The cyclic alternation of clay and sand reflected the periodic shifting of channels and the oscillating intensity of fluvial transport, superimposed upon the trend of gradually decreasing clastic inflow. The concentrated brines of the basin, first diluted during the deposition of the underlying anhydrite, became further diluted by the increased inflow of terrrestrial water; independent pollen studies have also reflected this temporarily risen humidity.

During this relatively humid interval, connections with the open sea severed during the preceding evaporite deposition were temporarily restored: the shale contains brachiopods, molluscs and marine algae. (Widening of the pseudomarine zone due to dilution may also account for the spread of life.)

Magnesite Unit

The carbonate layer overlying the Gray Shale consists of two magnesite beds separated by a central dolomite section. The basal magnesite bed is 6-33 cm thick and dark gray. This bed, locally pyritic, often contains several intercalating laminae and lenses of shale.

The central, dolomitic layer is lighter; it is characterized by rhythmic alternation of dolomite and magnesite laminae. Argillaceous laminae and irregular argillaceous patches occur rarely.

The upper magnesite bed is brown and contains small amounts of anhydrite and clay. Oolites, with ooids of 0.03 mm average diameter, occur locally. Discontinuous fissures, probably of desiccational origin, are often filled with anhydrite.

The total thickness of the Magnesite Unit is 0.8-2.25 m, increasing toward submarine shoals and toward the basin margin where it becomes gradually replaced by dolomite. This paleogeographic control of the magnesite, together with its interlamination with dolomite, indicates its primary to early diagenetic origin, possibly under the influence of heavy basin brines. An indication that the deeper basin areas may have been occupied by such hypersaline brines is provided by the observed restriction of the marine fauna to the basin margin and to some shoal areas. If this explanation is true, the rhythmic alternation of magnesite and dolomite might reflect periodic dilution of these heavier brines by terrestrial or oceanic inflow.

Alternately, if part of the overlying Main Anhydrite was formed in supratidal environment, the precipitation of $CaSO_4$ from the interstitial brines would sufficiently increase their Mg/Ca ratio to induce replacement of primary $CaCO_3$ or dolomite by dolomite or magnesite during early diagenesis. In this case, however, we would expect the magnesite content to decrease basinwards, while reported observations indicate its decrease in opposite, shoreward direction.

Main Anhydrite

The basal evaporite unit of the third Zechstein cycle, the Main Anhydrite, is 10-66 m thick. Cyclic bedding is distinct: thin beds of pure anhydrite, 5-10 cm thick, alternate with thin laminae (2-3 mm) of argillaceous-magnesitic anhydrite throughout the sequence; both the magnesite and the clay content decrease upwards as the anhydrite crystal size increases. The top 20-40 cm of the anhydrite suite consists of a shale bed, up to 20 cm thick, overlain by a similarly thick bed of anhydrite (Anhydrite Crust). These two apparently mark the start of a somewhat distinct cycle, which soon leads into halite deposition.

The different types of lamination, on which Kosmahl (1969) based the microstratigraphy of the Main Anhydrite, are briefly described below:

The lineated anhydrite is characterized by parallel, straight laminar planes. Anhydrite lamellae, 0.1-2.0 mm thick, alternate with argillaceous varves ranging in thickness from a few microns to 0.5 mm and averaging 30-50 microns. Magnesite crystals are rare.

The lamellar anhydrite is built up of undulating laminae;

both the anhydrite and the magnesitic-argillaceous laminae vary laterally in thickness, averaging 1-2 mm for the anhydrite and 0.5-1 mm for the magnesite laminae. Traces of anhydrite pseudomorphs after gypsum porphyroblasts are common, they are up to 2 cm large; transitions into flaser, schlieren, and porphyroblastic anhydrite occur.

In the banded anhydrite, the anhydrite laminae are very thick, 3-12 cm, and contain magnesitic flakes, schlieren, or dispersed magnesite. They are separated by magnesitic or argillaceous varves up to 5 mm thick, which consist of several distinct microlaminae. The bedding planes are wavy to even. Transitions into lamellar, flaser and porphyroblastic anhydrite occur.

In the flaser anhydrite, the wavy magnesitic laminae often split up into irregular wisps, which surround the often enterolithic anhydrite lenses. The anhydrite lenses themselves are frequently contaminated by magnesitic flakes, schlierens, and lamellae. Porphyroblasts up to 3 cm thick are common. Transitions into schlieren, banded, lamellar, flaky, and porphyroblastic anhydrite can be observed. The anhydrite lenses are up to 3 cm thick. The flaser zones are 3 mm to 5 cm thick, and consist of one or more magnesite ar argillaceous varves, 0.1-5 mm each.

In the flaky anhydrite, irregularly shaped magnesite aggregates, few mm in diameter, occur in the anhydrite. Transitions into schlieren, flaser, and porphyroblastic anhydrite can be observed.

In the porphyroblastic anhydrite, gypsum porphyroblasts, replaced by anhydrite, occur either as individuals in a carbonate matrix, or form disoriented aggregates in which the individual porphyroblasts are separated by magnesite or argillaceous network or laminae. The size of these pseudomorphs ranges from 1 mm to 5 cm. Transitions into schlieren, flaky, lamellar, banded and flaser anhydrite can locally be observed.

Thirteen zones, each having a characteristic type of lamination and each traceable through most of the Zechstein basin, have been differentiated by Kosmahl (1969). They can be grouped into seven cycles, each characterized by high magnesite (in the seventh cycle clay) content in its lower and by purer anhydrite in its upper portion. In the basal, magnesitic sections the magnesite laminae are both thicker and more closely packed than in the purer upper sections. The thirteen zones, marked by Greek letters, are presented below:

Cycle VII
- Omega: Lineated anhydrite (Anhydrite Crust). 0.08-4.3 m
  - Black Shale bed. 0.05-0.25 m

Cycle VI
- Lambda: Banded anhydrite. 2-12.8 m } Absent outside shoal areas
- Kappa: Porphyroblastic anhydrite. 0.7-13.2 m } Absent outside shoal areas
- Iota: Flaser, subordinately banded anhydrite. 0.1-7.7 m

Cycle V
- Theta: Alternating laminae of pure and flaser anhydrite. 4.4-26.5 m
- Eta: Lamellar anhydrite. 0.01-0.7 m

Cycle IV

Zeta: Alternating laminae of lamellar and schlieren or flaser anhydrite. 0.7-12.2 m

Cycle III

Epsilon: Schlieren anhydrite. 3-13.5 m

Cycle II

Delta: Flaser anhydrite with magnesite "nests". 1.6-3.3 m

Gamma: Lamellar anhydrite. 0.01-0.25 m

Cycle I

Beta: Flaky or flaser anhydrite, subordinately schlieren anhydrite. 0.3-1.4 m

Alpha: lamellar anhydrite. 0.4-2.2 m

More detailed analysis of the lamination shows that in the first five cycles the magnesite-rich lower portion consists of lamellar or schlieren anhydrite, in the seventh of shale, and only in one cycle, the sixth, of flaser anhydrite, whereas flaser anhydrite is common in the upper, magnesite-poor portion of most cycles. One aspect of the cyclicity is, therefore, that the precipitation of anhydrite or gypsum is more rapid in the shallow-water or supratidal environment where the development of the lenticular flaser-anhydrite is preferred.

The texture of the Anhydrite Crust is different from the rest of the sequence: its porphyroblastic section contains large amounts of anhydrite pseudomorphs after gypsum porphyroblasts, but most of the zone is dense to finely crystalline and finely and evenly laminated (lineated anhydrite). This major part of the thin Anhydrite Crust has, **possibly**, formed as primary anhydrite: it lacks the microscopic features which are common throughout the sequence and indicate secondary anhydrite formation from primary gypsum (like a certain succession of anhydrite generations, microcavities filled with sylvite, halite and carnallite, etc). Over the shoal areas, however, this primary, lineated anhydrite grades into flaser anhydrite exhibiting features characteristic of secondary anhydrite.

While the thickness of the Main Anhydrite is normally 25-40 m, local thickenings, rising over 35 m above the average top level, are common. These anhydrite "bars" occur over the shoals of the preceding Gray Shale times, indicating that rapid sulphate precipitation in shallow water was responsible for their formation. Although all zones of the Main Anhydrite become thicker in these anhydrite bars, the type of their lamination remains unchanged, indicating that the differences of the depositional facies were smaller than either during the formation of the basal anhydrites of the Stassfurt Cycle, or during the deposition of the Anhydrite Crust at the top of the Main Anhydrite.

Differences in the thickness of zones over shoal and deep areas generally correspond to similar changes in the thickness of the individual anhydrite laminae that build up the zones. The thickness differences are relatively less pronounced in the lower part of the Main Anhydrite, increase upwards, and become extreme in the two zones (Kappa and Lambda) which underlie the Black Shale bed: these two zones are thick in the shoal regions, but absent over the deeper basin areas. The overlying Black Shale bed also displays significant facies differences:

over some shoal areas it is replaced by porphyroblastic anhydrite containing only thin lenses and laminae of black shale.

Thus the gradually rising shoals increasingly accelerated sulphate sedimentation in their areas, leading ultimately to the complex basin topography that controlled the deposition of halite and potassium salts during the formation of the overlying Ronnenberg Group. Outside the shoals, the appearance of evenly laminated (lineated) anhydrite at the top of the anhydrite suite, in the upper portion of the Anhydrite Crust, may reflect the stabilization of basin-type environment preparatory to the start of halite deposition.

Fig. 15

Black Shale bed near top of light gray anhydrite, separated by the thin Anhydrite Crust from the non-laminated basal portion of the Ronnenberg salt. Height of section: 3 m.

Leine Cycle: contact of Main Anhydrite and Ronnenberg Group; Hansa Mine, Hannover, Germany.

Fig. 15

Ronnenberg Group

The Main Anhydrite is overlain (Fig. 15) by the Ronnenberg and Riedel Groups, composed of halite and potassium salts and forming together the salt sequence of the Leine Cycle.

The Ronnenberg Group is well exposed in the Hansa Mine near Ronnenberg in the Weser depression, and has been intensively studied by Hofrichter (1960) throughout the northwestern part of the German Zechstein. Here we shall follow his statements, supplemented by our personal observations in the Hansa Mine.

The Ronnenberg Group consists of a lower lineated and an upper banded salt, with the Ronnenberg potash deposit in between.

Lineated salt

In the Lineated salt, usually three zones are differentiated. The basal zone is gray to brownish red(Fig. 15); laminae of anhydritic halite, 1-4 cm thick, mark the indistinct lamination. Upwards the percentage of anhydrite in the salt decreases from 5 to 1 % as the color becomes paler.

In the middle section, the varves follow at remarkably constant intervals: 5-20 cm halite beds (Fig. 16) are separated by varves of anhydritic-kieseritic halite, 3-5, rarely 10 mm thick, in which anhydrite always predominates over kieserite. Anhydritic-kieseritic laminae of "second order", very indistinct, occur within the halite beds, along them and sometimes also along the major laminae the halite turns reddish, reminiscent of the internally laminated halite beds of the Werra Cycle (Fig. 6).

In tectonically disturbed areas, the laminae become indistinct or disappear, as the dark pigment of the anhydrite is oxidized. The colour of the halite is pale gray to pale brown; the anhydrite content of an average section is less than 0.5 %.

In the upper section of the Lineated Salt, lamination decreases. The anhydritic laminae are less distinct, their anhydrite is more dispersed, and the distance between them increases and becomes irregular.

A peculiar feature of this top zone, frequently orange-colored, is the presence of halite "augen", 1-2 cm in diameter. Thrir red tint is deeper than in the enclosing halite and they are frequently arranged in discontinuous laminae.

The method of varve-correlation, carried out by Hofrichter (1960) over large areas of the basin showed that the three zones of the Lineated Salt are not time-constant but are separated by time-diagonal boundaries.

The top zone is differentiated from the central one not only by its color but also by its potassium content: the K-concentration increases upwards. This increase is pulsating: the K-content abruptly drops at the top of each salt bed, to rise in the overlying one to a higher concentration than attained below. The highest K-concentration of the uppermost

<u>Fig. 16</u>

Pale orange halite beds, 25-60 cm thick, alternating with dark laminae of anhydritic halite.

Leine Cycle, Ronnenberg Group, Lineated Halite; Asse Mine, Braunschweig, Germany

Fig. 16

beds leads then over into the Ronnenberg potash deposit. The thickness of the entire salt formation between the Main Anhydrite and this potash zone is about 40-50 m.

Ronnenberg potash deposit

The potassium-bearing deposit that grades from the Lineated Salt is 2-6 m thick and consists mainly of white to gray to red sylvinite (sylvite+halite). Most potassium is contained in the lightly colored beds. The lamination is less pronounced: the anhydrite and kieserite, segregated into distinct laminae in the Lineated Salt, is here more or less evenly distributed. Locally, alternation of reddish brown and gray sylvinite beds of essentially identical composition can be observed.

Banded Salt

The halite between the Ronnenberg potash deposit and the Riedel Group is about 40-50 m thick, equal in thickness to the Lineated Salt. Here, too, three zones are differentiated.

The bottom zone, 10-12 m thick, consists of very pure, clear to pale gray or pale red halite. The varves, consisting of anhydrite with subordinate kieserite, are indistinct; the halite beds between them are thicker than those below the potash deposit and vary widely in thickness.

The middle zone is somewhat thicker, 12-15 m, and the cycles are slightly different. Twenty to forty reddish brown bands, about 3 cm thick each, alternate with pale halite beds, approximately 6 cm thick. The top of the zone is marked by the appearance of kieserite laminae. This kieseritic halite, widespread throughout the basin, indicates the return of high brine concentrations which eventually lead to the formation of the overlying local, small potash deposits of Habighorst and Bergmannssegen.

The top zone is the thickest one (18-23 m) and the most intensely colored; it consists of red and brownish red halite.

Riedel Group

Salt with anhydrite beds

The Riedel Group, the upper section of the Leine Cycle, commences with an anhydrite-rich zone. The anhydrite forms 8-10 interbeds in the halite, of which the fourth and the sixth are 1-2 m thick, while the rest are only 5-20 cm. The thicker anhydrite beds are argillaceous in their basal sections, and often contain halite in their upper portions, thus reflecting in their composition the progressive phase of evaporite cycles. Frequently, the anhydrite has a flaser texture: irregular anhydrite lenses, 2-7 cm in diameter, are separated by thin (1 mm) wisps and networks of argillaceous dolomite. This flaser anhydrite often passes laterally, even within the same mine, into anhydrite with large pseudomorphs after gypsum.

The halite zones between the anhydrite beds are 2-8 m, rarely 20 m thick, they constitute the majority of this 35-m formation.

Compared to the Lineated Salt of the underlying Ronnenberg Group, the palaeogeographic distribution of the anhydrite beds in this basal horizon of the Riedel Group shows some important differences. Thus, in the marginal areas, the basal section of the Lineated Salt contains considerable amounts of anhydrite; then, in the higher horizons of the Ronnenberg Group, the amount of anhydrite gradually decreases, until in this basal zone of the Riedel Group, several anhydrite beds that are present in the deeper basin areas are absent from the basin margin sequences. Inversely, in the deeper basin sections, anhydrite is relatively rare at the base of the Ronnenberg Group, while in the basal section of the Riedel Group all anhydrite beds are present. Hofrichter (1969) attributes this inversion of the paleogeographic pattern to a gradual decrease of marine sulphate supplies with progressive evaporation. According to him, at the start of evaporite sedimentation ample $CaSO_4$ reached and was preferentially precipitated in the warm shallow water of the marginal areas, while in later stages the $CaSO_4$ was essentially removed from the brine before it could reach the marginal regions.

Signs of reworking and short lasting withdrawal of the brine occurred first in the Main Anhydrite. Here such indications are common, expressed mostly in frequent channelling and suncracks. A local potash deposit overlying the seventh anhydrite interbed displays marks of dissolution and reprecipitation, indicating that at least part of the anhydrite interbeds may have been formed as a result of selective leaching of previously deposited anhydrite-bearing salt.

Swath Salt

Abovethe basal zone of the Riedel Group,the concentration of anhydrite further decreases in the halite. Thin anhydritic and sometimes argillaceous halite laminae, 1-3 mm thick, alternate with thicker salt laminae (1-5 cm). Intervals in which the anhydrite laminae are white and indistinct alternate with intervals wherethey are more numerous and distinct, stained gray by pyrite, bitumen and clay. These latter, more contaminated zones, which appear in the exposures as swaths of dark laminae, are called swath zones, and the entire horizon is named Swath Salt.

Herde (1953) subdivided the Swath Salt into five cycles or ten zones, the zones being alternately "pure" and "swath":

Cycle V			
	Zone 10.	Swath zone.	1-10 m
	Zone 9.	Pure zone.	1-20 m
Cycle IV			
	Zone 8.	Swath zone.	0.3-3.5 m
	Zone 7.	Pure zone.	0.5-4 m
Cycle III			
	Zone 6.	Swath zone.	1-10 m
	Zone 5.	Pure zone.	0.8-11 m
Cycle II			
	Zone 4.	Swath zone.	0.5-20 m
	Zone 3.	Pure zone.	0.8-7.5m
Cycle I			
	Zone 2.	Swath zone.	1-20 m
	Zone 1.	Pure zone.	0.2-4.5m

Within each cycle, the basal pure zone passes upwards gradually into the contaminated through a zone of "normal" salt with anhydrite laminae in the amounts most common in other halite zones of the German Zechstein. Within the swath zone, this increase of contamination continues to the top: the purest basal section of the next cycle overlies the most contaminated part of the swath zone without transition. Only the uppermost swath zone presents an exception: its most contaminated top zone grades, through a thin transitional zone of decreasing anhydrite and increasing potash content, into the Riedel potash deposit. A smaller, local potash deposit occurs above the eighth swath zone.

Comparison of this sequence with the underlying, basal member of the Riedel Group (Salt with anhydrite interbeds) reveals an important difference. While in the latter unit the anhydrite is most abundant at the bottom of each cycle, decreases gradually upwards, and then increases sharply to form the next interbed, in the Swath Salt the asymmetry is inverse: At the bottom of each cycle the salt is pure, further up dark-stained anhydrite laminae intercalate; contamination and anhydrite content increase gradually in the overlying swath and cuminate at its top only to drop abruptly to nil in the pure halite at the base of the next cycle. Herde (1953) attributes this inverse asymmetry to leaching; we shall return to this interpretation in the chapter discussing genetical problems.

Riedel potash deposit

The major potash deposit of the Riedel Group, which overlies the uppermost contaminated zone of the Swath Salt, is normally 2-6 m thick. The shallow brine depths and diversified bottom relief have led to great facial diversity: apart from a gray and a marginal red sylvinite facies there are also a carnallite and a white halite facies. Intercalation and wedging out of the sylvinite layers in low-potassium sylvitic halite is common, particularly in the gray sylvinite facies. Scuh lateral changes are apparently primary phenomena: high-grade sylvinite was deposited in deeper pools that collected the more concentrated heavy brines, while over the shoals in between low-grade sylvitic halite formed. On the other hand, late diagenetic leaching was probably responsible for the formation of the sylvite-free white halite beds.

The appearance of gray and black clay flakes marks the impending change in sedimentation; in marginal areas both the Swath Salt and the overlying Riedel Deposit are argillaceous, the latter normally red. At the top of the Riedel Deposit clay becomes generally more abundant, and this argillaceous bed is overlain with a sharp contact by a sylvitic bed containing corroded grains of halite and sylvite. The process of leaching and re-precipitation presumably took place in local shoals and marginal areas.

Temporary emergence and selective dissolution may account for the fact that the area occupied by the discontinuous Riedel deposit is much smaller than the one occupied by the continuous Stassfurt potash deposit of the second Zechstein cycle; while the centers of the two deposits more or less coincide (Fig. 17) the outer limit of the Riedel deposit is about halfway between this center and the limit of the Stassfurt deposit.

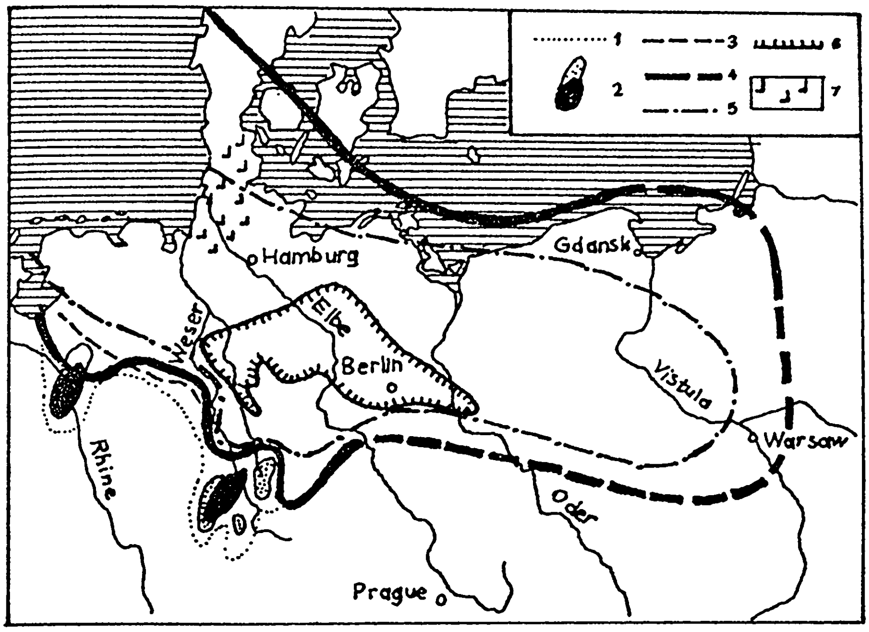


Fig. 17. Distribution of salts in the Permian of the North German Basin (Fiveg, 1957). 1. Southern limit of Werra salts; 2. Werra salts more than 100 m thick, with area of potash deposits blackened; 3. Northern limit of Werra salts; 4. Limit of Stassfurt salts; 5. Limit of the Stassfurt potash horizon; 6. Limit of the Leine potash horizons; 7. Lacustrine salt in the Lower Permian.

Salt with shale laminae

Both the leached halite facies of the Riedel deposit and the overlying bed that contains corroded grains of sylvite and halite witness increasing leaching. Parallel increase of clay content leads over from the Riedel deposit into a zone of halite, 20 m thick, containing shale laminae. As in the axial zone of the Appalachian basin (Part II), the shale occurs in the form of intercrystalline greenish gray to red "fragments", 2-10 mm large, or forms laminae of similar thickness. The "fragments" seldom display any internal lamination, and do not indicate reworking. Their shape is controlled by the shape of the halite crystals among which they occur. According to Herde (1953), this structure indicates deposition of minor amounts of argillaceous mud on a bottom covered by angular halite crystals; the mud has filled their interstices but, in most instances, failed to cover them completely. In our opinion, recrystallization after sedimentation played a major role in the formation of this structure.

Upwards these fragments and flakes become increasingly abundant, until they eventually merge into continuous laminae at the top of the first cycle of this shaly salt. The average thickness of this first cycle is 5.75 m.

In the area around Ronnenberg, However, salt is absent from the first cycle: it is replaced by 3 to 5 m of bluish black shale, the so-called Blue Shale. The shale is well laminated, its laminae, 2-6 cm thick, are intensely pyritic.

Yellow koenenite frequently occurs between the shale laminae or between shale and halite. Alternation of anhydritic and anhydrite-free shale laminae may reflect seasonal changes in brine concentration.

According to Herde (1953), the presence of koenenite indicates that the sediments of the first cycle deposited or redeposited in terrestrial water, carrying muds, and washing in previously deposited and now subaerially exposed salt into the basin. In his view, no appreciable quantity of marine brines entered the basin during this period.

The Blue Shale, according to him, deposited in a marginal sub-basin, which served as a settling tank eliminating the suspended solids from the inflowing brines.

The top of the uppermost shale interbed, or, where the halite is absent, the top of the Blue Shale, is extremely sharp, suggesting a break in sedimentation similar to those at the top of the contaminated zones of the Swath Salt. Apparently, here as well as in the Swath Salt, this break is the result of intraformational leaching leading to the accumulation of the less soluble components.

The second cycle of the Salt with shale laminae starts with pure, clay-free salts, overlying the sharp top contact of the first cycle. In the same marginal depression where the blue Shale was deposited during the first cycle, an areally restricted potash deposit occurs in this horizon: the Albert deposit. As the underlying local deposits of sylvinite, the

Albert deposit, too, shows signs of redeposition: corroded and rounded halite crystals are common. Apparently, highly soluble salts, first deposited over a wider area, were ultimately washed into the same margianl depression which, during the first cycle, settled the clastics carried into the basin.

The Albert potash deposit is overlain by halite which is still part of the second cycle and contains flakes, fragments and laminae of shale in upward increasing quantities. The uppermost shale bed is accompanied by an anhydrite lamina, perhaps indicating new inflow from the open sea. The average thickness of the second cycle, excepting the local potash deposit, is 3.40 m. the top surface of the cycle is again very sharp.

The third cycle is only 1.75 m thick; its structure resembles the second cycle with upward increasing shale content, leading into shale laminae with upward increasing thickness. The uppermost of these is again accompanied by an anhydrite lamina which is overlain by the red Boundary Salt that marks the top of the Leine Cycle.

Otherwise, as in the axial zone of the Appalachian basin (Part II), anhydrite is conspicuous by its absence in the Salt with shale laminae. According to Herde, this indicates very little inflow from the open sea, and a hydrochemical system controlled primarily by the inflow of terrestrial water dissolving and redepositing previously deposited marine salts. Within each of these cycles of reworking, the bottom consists of pure evaporites, while signs of leaching and laminae of shale become increasingly frequent upwards. In other words, very short periods of progressive evaporite sedimentation alternate with very long recessive (decreasingly evaporitic) periods, the latter lacking the $CaSO_4$ stage.

Z4 - ALLER CYCLE

Red Shale

With upward increasing clay content, the Leine salts gradually pass into purely clastic sediments. Flakes and fragments of gray shale become increasingly common in the Boundary Salt at the top of the Leine Cycle, and they soon merge into a greenish gray, rarely bluish black shale layer. This shale layer is taken as the base of the uppermost major Zechstein cycle, the Aller Cycle. This "Green Foot" of the Red Shale is laminated, and contains small shale balls in large amounts.

According to Herde (1953), this shale represents the least evaporitic top of the uppermost cycle of the Salt with shale laminae, and its high clastic content is, at least in part, the result of temporary leaching preceding the deposition of the overlying halite beds. Like the top of the underlying cycles affected by such dissolution, the top of the "Green Foot" is very sharp.

This sharp top contact is overlain by the relatively pure halite base of the subsequent cycle. Flakes and fragments of shale, rare at the bottom, become increasingly frequent toward the top, while their color turns from grey into red. Eventually, they merge to form the Red Shale proper, which is the terminating part of this second cycle. On a larger scale, this shale represents the culmination of the trend from salt to shale, manifest in the upper part of the Leine Cycle.

The Red Shale often contains sand and silt, as well as some carbonate. The major clay minerals are illite and chlorite, reflecting the high ion concentration of interstitial brines both during sedimentation and during diagenesis. (The clay minerals of the shales in the German Carboniferous consist, to a large extent, of kaolin and illite, reflecting the intensive leaching of inter-layer cations characteristic of in humid climate). Red halite nodules, 5x20 cm large, are particularly common in the basal section of the Red Shale; haematite flakes are common throughout.

The thickness of the entire Red Shale, including the "Green Foot", is about 20 m.

Pegmatite Anhydrite

The basal anhydrite of the fourth major Zechstein cycle is reddish gray at the bottom, gray to white at the top. Hypidiomorphic pseudomorphs of red salt after gypsum give the anhydrite a pegmatite-like structure. The halite crystals are 5-10 mm large.

Locally, the lower red and the upper white horizons are distinctly differentiable; in the red portion, large nodules and veinlets of red halite are common, while the upper, white portion contains only microscopic halite. The average thickness of the entire anhydrite layer is 1.5 m.

The color change from red to white or grey refelcts decreasing oxygen content of the interstitial brines, probably due to increasing flooding by, or increasing depth of, poorly circulating basin brines of low oxygen content. This change in the depositional environment introduces the massive precipitation of halite.

Halite with anhydrite laminae

The anhydritic portion of the Aller halite suite is 45 m thick, it consists of relatively pure salt. Three, in part probably time-equivalent, sections are distinguished.

In the basal salt, directly above the Pegmatite Anhydrite, the halite beds are pale pink, 6-30 cm thick. They are separated by slightly kieseritic anhydritic salt laminae, 1-5 mm thick; in these laminae, kieserite comprises less than 1/15 of the anhydrite content.

The basal salt gradually loses its anhydrite content, so that the cyclic lamination becomes indistinct. In this clear, frequently white salt, the NaCl-content is normally above 98%. Anhydrite and some kieserite, in the same proportions as below, are only minor contaminants.

The top pink salt forms a transition toward the overlying argillaceous salt. Though the lamination is still indistinct, white anhydrite laminae occasionally can be distinguished. Near the top, they are replaced by laminae of brown shale.

Salt with shale laminae

As the Leine Cycle, the Aller Cycle becomes increasingly argillaceous in its upper portion. In the pink salt, laminae of greenish-gray shale are 0.5-2 cm thick. Flattened shale balls, 1-5 cm, frequently occur in these shale laminae; often, they are thicker than the laminae themselves and protrude from them into both the underlying and the overlying halite bed. Near the top, these shale balls turn red and reach 10 cm in diameter; they contain red halite nodules. Apparently, they were formed through reworking of red shales similar to those at the base of the Aller Cycle.

The entire Salt with shale laminae is 55 m thick; it consists of approximately 96-98 % halite, 1.5 % anhydrite, and 0.5-2 % magnesite and clay, the latter increasing upwards. (Dr. Bauer, oral comm., 1970). Several features of it, like the presence of shale balls,interstitial shale network and shale laminae in the halite, the scarcity of anhydrite, and the presence of halite nodules in the shale, resemble the similarly shaly upper Silurian salt in the axial zone of the Appalachian basin, deposited close to the Taconic mountain-islands (Part II). In the Zechstein, these features may indicate a resurgence of tectonic activity in the Variscan hinterland. The increasing tectonic activity, resulting in increasing inflow of clastics into the basin, soon terminated evaporite deposition: in the overlying Lower Triassic, red shales, deposited mostly in terrestrial environments, become dominant.

FACIES, GENESIS AND CYCLICITY

Shales

As discussed in the introductory section, the clastic sediments form detritic wedges along the basin margin, thinning from the Variscan framework toward the basin interior, in the Zechstein as well as in the underlying and overlying detritic sequences (Rotliegende and Buntsandstein). Near the margin, all shale beds are red, indicating that more oxygen was contained in the interstitial brines of the sediment than what was needed to oxidize the organic matter during diagenesis. Basinwards, gray and black, often fetid shales become common, particularly in the lower Zechstein cycles. (In the uppermost, Aller Cycle, the shale maintains its red color throughout the entire basin.) The dark shales do not necessarily imply greater depths of deposition, but rather signify permanent flooding by organic-rich stagnant brines. Thus small depths of deposition are witnessed by the wedging out of lower copper shale horizons along the flanks of the underlying dunes (Fig. 4).

Lamination is common, particularly in the gray shales; sandy or silty laminae, 1-30 mm thick, intercalate at intervals often considerably exceeding their thickness. In most shale layers, the grain-size and amount of the sand decrease upwards, reflecting the connection of the shales with upward fining fluvial sequences. The alternation of coarser and finer laminae reflect the often seasonally oscillating precipitation in the Variscan mountains, as well as minor shifts in the position of river channels carrying the detritus. The correlation of shale layers with more humid periods has been confirmed by pollen studies.

With the decreasing grain-size, the carbonate content often increases upwards, as in the copper shale of the first or the gray shale of the third cycle. The carbonate minerals are generally dolomite or magnesite, indicative of deposition or replacement in highly saline environments. Possibly, the observed precipitation of $CaSO_4$ in discontinuous (desiccational?) cracks and in the interstices of the shale increased the Mg/Ca ratio of the interstitial brine (Kinsman, 1966, 1969), creating chemical micro-environments favorable for metasomatic dolomite or magnesite formation.

Halite is a common component of all shales in the German Zechstein. In most cases it is finely and evenly dispersed, but in some of the red shales it forms more or less isometrical red nodules. Part of these nodules formed, as in the shales of the Appalachian Salina sequence, as concretions in the desiccating argillaceous mud, others may be the relics of halite beds leached during early diagenesis.

Thicker shale beds often accumulate as a result of prolonged intradepositional leaching of halite layers containing shale laminae. In such cases, the top portion of the shale may have deposited as non-halitic muds from relatively dilute terrestrial brines, which simultaneously dissolved downwards decreasing quantities of halite from the underlying sediment. As a result, the shale laminae in the halite underlying the shale become increasingly closely spaced upwards, and they quasi merge into the overlying shale layer. This shale layer, then, is overlain by relatively pure halite, marking the base of a new cycle of evaporite deposition.

Carbonates

The carbonates, grading upwards from the basal shales, often display a characteristic facies pattern. The basin margin is lined by thick, bioconstructed, bioclastic and oolitic carbonates, in general light-colored and massive; the interior is covered by a thin carbonate sheet, argillaceous-bituminous and finely laminated. This pattern indicates well-ventilated, agitated water along the basin margin, and stagnant water in the interior. The traditional interpretation of salt genesis assumes that the very thick biogenic and bioclastic carbonates at the basin margin were formed contemporaneously with the thin fetid laminated carbonate sheet of the interior, and that the growth of reefs, increasing the isolation of the interior, were instrumental in the salt deposition. In this case, the height of the reefs would reflect the depth of the basin that later received and was rapidly filled by the salts. This correlation is, however, imperfect: the salt deposits of the first cycle occur in marginal position, those of the second cycle transgress beyond the underlying fetid laminites, and in the fourth cycle no major carbonate bed occurs.

The carbonates often display considerable sulphate and magnesite content. The sulphates occur in the form of (often leached) concretions close to the margin; the magnesite content either increases (second cycle) or decreases (third cycle) from the margin toward the interior. Increase toward depth may be due to density stratification of the brine; increase toward the margin may result from subaerial sulphate precipitation and concomitant changes in the composition of interstitial brines, as in the shales discussed above.

The interlamination of magnesite and dolomite in somewhat oolitic carbonates (third cycle) witnesses syngenetic or very early diagenetic magnesite formation. The concomitant alternation of argillaceous and clay-free laminae may indicate correlation of these small cycles with oscillation in the detritus inflow into the basin. These small cycles might be due to the same cycles of oscillating humidity in the mountainous hinterland which had controlled the oscillating grain-size of the underlying shales.

Anhydrite and gypsum

General features

In the lower two cycles of the Zechstein, the distribution of sulphate rocks follows the facies pattern of the underlying carbonates: thick, light, often massive sulphates occur along the basin margin, while a very thin sheet of fetid laminated sulphates comprising less than 1% of the marginal sulphates (0.8 m in the second cycle) covers the interior. Even single laminae of these sulphate laminites show excellent correlation over wide areas.

In the third cycle, the fetid laminites are insignificant (they line only the top of the anhydrite suite); in the fourth cycle they are entirely absent.

Apart from these evenly laminated ("lineated") sulphates, there are several other varieties characterized by increasing irregularity of the lamination. At the other end of the spectrum is the flaser anhydrite, which displays a lenticular-enterolithic structure.

Most structural features of the sulphate rocks are marked by laminae or (in the flaser anhydrite) wisps of carbonates; the carbonate mineral is dolomite and, more often, magnesite, suggesting the same causal relationship between magnesite and sulphate genesis as indicated before. Where evenly laminated and flaser anhydrites occur in the same horizon, as in the first and second cycles, the latter increase in proportion toward, and often become exclusive in, the marginal and shoal areas. The indistinctly laminated varieties (schlieren, banded, flaky anhydrite) of the third cycle show no such relationship to the flaser anhydrite.

Primary deposition of gypsum is often indicated by the presence of pseudomorphs after gypsum porphyroblasts, microcavity systems filled later by the most soluble salts, and a certain sequence of anhydrite generations (Kosmahl, 1969).

Locally, as in the pegmatite anhydrite of the fourth cycle) the gypsum porphyroblasts are replaced by red halite.

The genetical problems of flaser anhydrite

German authors generally assume that in the semi-laminar sulphate varieties (schlieren and flaky anhydrite) $CaSO_4$ and carbonate or clay were deposited simultaneously, though in oscillating proportions. Their even or at most slightly wavy structure is attributed to deposition in a relatively quiet sedimentary environment.

The problem of the lenticular flaser anhydrite is more complex. In the first and second Zechstein cycles, they were preferentially deposited along the basin margin, apparently contemporaneously with, but several times exceeding the thickness of, the laminated anhydrites of the basin interior. This greater thickness has been attributed to quicker deposition in the warmer shallow water, an approach supported by the thickness pattern of the Main Anhydrite of the third cycle, in which the thickness maxima of each bed occur over the same shoals, progressively increasing their height. Mere rapid deposition is, however, hardly sufficient to account for the lenticular, flaser structure itself. Several schools of thought have attempted to find an answer to this problem (Kosmahl, 1969):

According to the school of primary genesis (Herrmann and Richter-Bernburg, 1955), the lenticular anhydrite was formed in agitated water. This theory envisages disturbed sedimentation during most of the deposition of the Main Anhydrite, alternating with conditions of quiet sedimentation in calm water. This idea was further developed by Langbein (1961, 1969) on the basis of microscopic investigations. Langbein, finding good correlation between the palaeogeographic location of a given anhydrite profile and the micro- and macroscopic structure of the anhydrite, concluded that the latter must be faciologically controlled. According to him, this "sphaerolitic" structure was already present in the primary gypsum, and escaped destruction during the transformation of gypsum into anhydrite. He attributes this structure to $CaSO_4$-precipitation from low-salinity brines, which would explain why they grade towards the basin center as well as upwards (toward the overlying halite deposits) into so-called granoblastic or porphyroblastic structures attributed to higher salinity.

The school of early diagenetic formation of flaser anhydrite assumes (Jung, 1958, 1960) that all sulphates were originallyevenly laminated, and all other anhydrite structures developed diagenetically from this prototype.

Though it would be interesting to apply here the information gained by Shearman (1963, 1966) and Kinsman (1966, 1969) from studies in the Persian Gulf area, to my knowledge no such attempt has yet been made. The preferential occurrence of lenticular anhydrite in near-shore areas, its regular alternation with laminated anhydrite varieties, and the prevalence of the laminated anhydrite in the vicinity of the overlying halite indicate that such an approach may be promising. Thus the dominance of semi-laminar and flaser varieties in the Main Anhydrite may be due to frequent emergences after the deposition of the underlying Magnesite Unit, and the appearance of lineated anhydrite at the top, directly below the overlying halite, might indicate renewed submergence below more concentrated brines. The great differences between the lenticular anhydrites of the basin margin and the evenly laminated anhydrites of the basin interior in the lower Zechstein cycles as well as in the Anhydrite Crust at the top of the Main Anhydrite might be better accounted for if one assumes that they were formed in substantially different (subaqueous vs. supratidal) environments, than if one has to explain them with differences in brine temperature and salinity.

On the other hand, the studies of Richter-Bernburg convincingly indicate that the laminae of the lineated anhydrite were formed as continuous sheets covering the entire basin rather than as linear structures in a migrating tidal zone. Perhaps the contrast between the Persian Gulf of almost normal marine water, and the Zechstein basin filled with brines saturated with $CaSO_4$ may account for this divergence. While in the Persian Gulf area subaqueous evaporites can form only in the intertidal position, in the Zechstein basin they could form subtidally as well, over the area of the entire basin, as the overlying halite layers did. The existence of such a large hypersaline reservoir would also accelerate the precipitation of supratidal flaser sulphates along its shores.

Early diagenetic transformation of gypsum into anhydrite may also have affected the sedimentary structures. D'Ans and Kühn (1960), Braitsch (1962) and Kosmahl (1969) assume that this process started with the formation of anhydrite nuclei in and on the surface of the gypsum sediment, or at the contact of primary gypsum and primary anhydrite beds, and spreading in a rapidly progressing chain reaction, altered the metastable gypsum into stable anhydrite. The anhydrite nuclei could form in shallow water above shoals and near the margin (D'Ans and Kühn, 1960) where the temperature, or in the basin (Richter-Bernburg, oral comm., in Kosmahl, 1969) where the concentration was higher and therefore favorable for anhydrite formation. The ensuing dehydration may have proceeded horizontally as well as downwards into previously deposited sulphates. Since the Anhydrite Crust is not disturbed by the escaping water, probably the dehydration of the underlying bulk of the Main Anhydrite took place before its deposition. The dehydration of the Anhydrite Crust, in its turn, perhaps starting from the interfaces of its primary anhydrite and primary gypsum layers, also took place before the deposition of the overlying halites, postponing their deposition by diluting the brine with the escaping water (Braitsch, 1962).

These dehydration processes alone would, however, not disturb the original structures sufficiently (Stewart, 1953; oral comm. 1971) to cause a general transformation of even lamination into flaser structure, as thought by some of the above authors.

Exponents of <u>late and post-diagenetic</u> control of the flaser structures, like Borchert (1954, 1959, 1963), Lotze (1957), Storck (1954), Kokorsch (1960), and Siemeister (1961) also attribute their enterolithic-microfolded character to the dehydration of gypsum, but assume that this process did not take place until the gypsum was covered by considerable overburden, in the lower Triassic or perhaps only in the late Mesozoic. The anhydrite floating in the newly escaped water was, according to this concept, highly responsive to tectonic pressure, resulting in diverse internal structural deformations. The shortcoming of this theory is that it cannot account for the absence of intensive leaching such a huge amount of water would perform in the overlying salt. Though the red tint, higher crystallinity, narrower lamination and smaller thickness of the Leine halite along the basin margin might be attributed to this water (though they might as readily be explained by other primary sedimentary and diagenetic processes as well), the lack of leaching along the halite-anhydrite boundary, where it should be most intensive, is a strong argument against such an interpretation. (Kosmahl, 1969).

Small anhydrite cycles

According to Richter-Bernburg (1955 and later), the alternation of anhydrite with carbonate or shale laminae indicates regular oscillation in the concentration of the brine from which they are deposited. He considers these sediments as varvites of annual layering, anhydrite or gypsum depositing in the warm season and carbonate or clay in the cold one. Studies by Richter-Bernburg in the basal anhydrite of the Stassfurt Cycle gave 0.2-1.2 mm (in average 0.6-0.8 mm) as the thickness of the annual anhydrite laminae, which agrees well with the estimated rates of annual evaporation. Richter-Bernburg and Herrmann (1955) regard, however, even the very thick (1-5 cm) laminae of the anhydrite as annual precipitates, the anomalous thickness of which was caused by anomalously high summer temperatures of a given year exceeding a critical threshold. Where such abnormally thick (or abnormally thin) laminae occur at 11-lamina intervals, the anomaly is attributed to the influence of sunspot cycles.

Braitsch (1962) points out, however, that the deposition of 1 cm anhydrite would require the evaporation of 8 m water, already evaporated to the point of $CaSO_4$-saturation, while the maximum known rate of evaporation for larger amounts of water is 2 m/year. This observation seems to indicate that laminae thicker than 1 mm were deposited over intervals of several years.

The Main Anhydrite of the Leine Cycle (Kosmahl, 1969) consists of cycles, a few m thick each, of alternating magnesite-rich and relatively pure Ca-sulphate. The magnesite-rich lower portions often display semi-laminar (banded, flaky, or schlieren) structures, while the magnesite-poor upper portions generally consist of lenticular, flaser anhydrite. Throughout the Main Anhydrite, there is a trend of gradual decrease in the thickness and magnesite content of the lower, magnesite-rich portion of each successive cycle. According to Seidel (1961) and Kosmahl (1969), this gradual increase in $CaSO_4$-purity, interrupted by the deposition of the Black Shale bed, reflects a gradual increase in salinity, eventually leading to halite deposition. Within each cycle, the progressive phase of increasing salinity is longer than the recessive one, resulting in rather abrupt appearance of magnesitic laminae over pure anhydrite. This asymmetric oscillation is further modulated by smaller cycles of local significance and by the still smaller cycles of presumably annual lamination.

Kosmahl (1969) assumes that these cycles are controlled by climatic changes. Preferred precipitation of the less soluble magnesite in the lower, and of the more soluble $CaSO_4$ in the upper portion of each cycle is caused by the increasing concentration of brines or, more accurately, by the decreasing solubility of $CaSO_4$ at increasing brine temperatures. Temperature differences between shallower and deeper brines may have given rise to current systems, to which Kosmahl tentatively attributes the flaser structure typical of the upper portion of each cycle.

These oscillations of climate were shortlasting and of relatively low amplitude, never leading to the deposition of pure halite. The general climatic trend was, according to Kosmahl (1969), a gradual transition from the more humid climate prevalent during the deposition of the underlying Gray Shale, through a semi-arid climate when the Main Anhydrite was deposited, into the more arid climate of the subsequent halite deposition. The increasing aridity of the mountainous hinterland was interrupted by a short period of relatively higher humidity, responsible for the deposition of the Black Shale bed near the top of the anhydrite suite. This intermission did not, however, substantially affect the concentration of the brines; after aridity had been reestablished in the hinterland brine temperatures had risen to their previous level, calcium sulphate deposition continued and gave place soon to halite precipitation.

After climate, the second important factor governing cyclic sedimentation may have been the oscillating level of the "bar" which separated the Zechstein Basin from the open sea to the north. Whether one visualizes this bar in the classical sense or as a saturation shelf, and whether these late Variscan movements were epeirogenic or orogenic, evidently even minor changes in the water depths of this region must have exerted great influence upon brine concentrations and temperatures in the basin. Erosional nonconformities at the base of many small cycles, underlain by flaser and overlain by laminar anhydrite, may have resulted from dilution due to the lowering of the bar level, although they might as well be considered contacts between supratidal and overlying intertidal evaporites by investigators favoring this mode of evaporite formation. Similarly, the gradual decrease of carbonates in successive anhydrite cycles may also be due to uplift or decreasing depth in the bar region. The specific evaporation of the resulting shallow water would be high, precipitating most carbonates, so that the brine entering the basin would already be depleted in carbonates.

Several aspects of the cyclicity may have been controlled by purely <u>sedimentological factors</u>. Such factors were responsible for the increasing differentiation of shoal areas as a result of more rapid sulphate deposition in their shallow water. The complete filling of shallow basin areas may have created supratidal flats favourable to flaser anhydrite formation. Cycles of submergence (as a result of locally uncompensated subsidence), filling and desiccation could produce cycles of subaquaeous laminated and supratidal flaser sulphates, separated by unconformity surfaces, without the necessity to invoke any climatic or tectonic change. Such cycles would be expected to become decreasingly distinct toward deeper basin regions, as indeed they are. The oscillating and decreasing occurrence of carbonates imply, however, a more distant lever of control: periodic entrance from outside the basin in gradually decreasing quantities. The fact that anhydrite horizons near the Black Shale bed, which evidently is a product of terrestrial inflow, are not characterized by larger quantities of magnesite, seems to indicate that the carbonate was imported by marine rather than terrestrial waters. In this case, its fluctuating and decreasing concentration in the sediments may also reflect oscillating inflow and gradually increasing isolation from the open sea, presumably due to oscillating and gradually decreasing depths in the bar area. These changes need not be attributed to tectonic movements: they may have been caused by fluctuating rates of sedimentation in the bar area, in a regime of continuous subsidence. These fluctuating rates may be climatically controlled: increasing aridity would produce more rapid precipitation from the brine. Thus the gradual decrease of detritic sediments in the Gray Shale, their disappearance in the Main Anhydrite, and their cursory reappearance in its Black Shale bed, as well as the presumably annual carbonate and shale laminae of the anhydrite, are all climatically controlled phenomena. The fact that these ultimately climatically controlled cycles consist of the same sequence of lithologies as the major cycles may indicate that the latter, too, could be caused by ultimately climatic oscillations.

Beyond the climatic, tectonic, and depositional factors, reworking is a major agent controlling cyclicity in the

Main Anhydrite. It is reflected by large-scale unconformities as well as small channels and pockets, occurring particularly between the first and second, within and on the top of the fourth, and between the sixth and the seventh cycle. Apparently, (Kosmahl, 1969), the channels and pockets are produced by currents, eroding and partially dissolving the gypsum.

The more extensive unconformities result from extensive leaching caused by substantial decrease in brine concentration. This is particularly evident in the case of the anhydrite underlying the Black Shale bed (seventh cycle): here the inflow of large amounts of terrestrial water into the basin led to the dissolution of previously deposited sulphates over extensive areas. Conceivably, the lack of the entire upper portion of the sixth cycle, underlying the Black Shale, in the deeper basin areas, is due to this dissolution. Even this leaching was, however, not equally effective over the entire basin. In local shoal facies (Grasleben, Beienrode), the anhydrite of the sixth cycle grades without hiatus into the seventh cycle, the intervening Black Shale bed being replaced by argillaceous porphyroblastic anhydrite which grades laterally into shale. This local absence of leaching, accompanied by a drop in the amounts of argillaceous material and by the continued precipitation of sulphates, may indicate that terrestrial water failed to reach these areas in substantial quantities.(Kosmahl, 1969).

Most of the gypsum, removed mechanically or chemically from previously precipitated sediments, was again redeposited in the same or in the succeeding cycle, only a small fraction, if any, being carried from the basin back into the open sea. Most of the redeposition took place, presumably, as soon as a renewed increase in the brine temperature decreased the solubility of $CaSO_4$ to the previous level, or the water that had caused the leaching evaporated. Thus most of the Anhydrite Crust, deposited after the Black Shale bed and before the halite, may have been derived from the gypsum leached from the sediment when the Black Shale bed was being deposited (Kosmahl, 1969). If the Anhydrite Crust was indeed deposited over the deeper areas as primary anhydrite, then this could reflect the higher brine concentrations that preceded the deposition of halite.

The reasons for the higher frequency of reworking during the third Zechstein (Leine) cycle are not clear. The argument of Kosmahl (1969), that the Zechstein Basin had been much deeper during the first Zechstein cycle and became gradually shallower and thus more favorable for reworking during the second and particularly the third Zechstein cycles, is open to discussion: there is not sufficient evidence that the basically similar argillaceous, sulfate and chloride facies, recurring in the same order in each major Zechstein cycle, were deposited at so fundamentally different depths. Whatever the cause, reworking here intensely modifies the control exerted over cyclic sedimentation by climate, tectonism and sediment accumulation.

Anhydritic halite, normal cycles

By far the majority of the halite is well bedded. The bedding is marked by anhydrite laminae, which are only a few mm, or fractions of a mm thick. Suites of regularly bedded salt, in which the anhydrite laminae follow at relatively constant intervals ("lineated salt") are common, but often the intervals are less regular, and the anhydrite laminae are replaced by somewhat thicker and less distinct laminae of halitic anhydrite or anhydritic halite.

The great thickness of the salt beds (8-20 cm) conceived of as annual deposits, was an important factor in interpreting salt basins as deep basins, rapidly filled with evaporites whose rate of deposition was several orders of magnitude higher than most known rates of subsidence. This interpretation has been altered by recent observations. Thus it has been found (Jung, 1959) that where the halites of the Werra Cycle grade laterally into anhydrite, the varves in the anhydrite are an order of magnitude higher than the number of the major anhydrite laminae in the equivalent salt section. Further research revealed sets of very thin anhydrite laminae, observable only in transparent light, between the more distinct, thicker anhydrite laminae separating the salt beds. In intervals where the salt is more pigmented (see discussion of the Werra halite), this internal lamination is more distinct. The sum of the thicker and thinner anhydrite laminae agrees well with the number of anhydrite varves in the lateral anhydrite equivalent (Jung, 1968).

Similar observations in the Stassfurt halite led Jung (1968) to conclude that the observed gross lamination in the halite reflects decennial rather than annual intervals, and that the thickness of the annual halite laminae is only 0.5-1.0 cm. This rate of deposition is, according to him, in good agreement with a projected regime of synsedimentary (rather than presedimentary) subsidence and "presents a further argument in favor of the general shallowness of the Zechstein basin".

The clay content of the anhydrite laminae indicates, as several bromine studies did before, that their deposition was controlled by periodic rainfall in the mountainous hinterland (compare Kunasz, 1970, concerning similar lamination in the Silurian salt of the Michigan Basin). If the most distinct, argillaceous anhydrite laminae are not annual, but decennial-undecennial deposits as Jung (1968) reports, it would indicate that the rainfalls occurred mostly at intervals corresponding to the present sunspot cycles. Such a great difference between the annual and the undecennial deposits seems to be considerably more than the present, indistinct difference between sunspot-rich and sunspot-poor years seems to warrant (Duff et al., 1967).

In normal anhydritic halite suites, the thickness and frequency of anhydrite beds decreases upwards with increasing distances from the basal anhydrite. This decrease is thought to reflect the gradually increasing concentration of basin brines, leading eventually to conditions favorable to potassium salt deposition.

Anhydritic halite, leached cycles

In halite suites containing anhydrite interbeds thicker thant the anhydrite laminae discussed aboe, the anhydrite is generally most abundant at the base of each cycle. The thickness of the intercalating anhydrite laminae, and with it the anhydrite/halite ratio, gradually decrease upwards, to increase again rather abruptly as the next thick anhydrite bed appears. In the leached halites of the Swath Salts (Leine Cycle; Herde, 1953) the asymmetry is inverse: At the bottom of each cycle, the salt is pure; upwards, contaminated, anhydritic laminae intercalate. Contamination and anhydrite content increase gradually in the overlying "swath", to drop abruptly to nil in the pure halite forming the base of the next cycle.

This upward increase in anhydrite content within each of the swath cycles is in conspicuous contrast with the general trend of gradual decline in the amount of anhydrite contained in each successive swath cycle. This decline, continuing the same trend from underlying normal cycles, reflects the gradual increase in brine concentration which leads ultimately to potash deposition.

Thus the main trend of progressive evaporite sedimentation, leading from anhydrite to potassium salts, is modified by two groups of factors controlling cyclicity. The first group contains the causative agents of cyclic sedimentation, like periodic changes in climate, tectonism and sediment accumulation. The second group comprises processes acting on previously deposited sediments and affecting their primary cyclicity.

These latter processes consist primarily in partial dissolution. From year to year, sediments of the upper part of each swath cycle, the swath zone, were subjected to increasingly frequent leaching, removing part of their halite and thus enriching them in anhydrite. Subsequently, physico-chemical conditions conducive to salt precipitation were reestablished, and the excess of the leached salts was rapidly redeposited as pure salt over the dissolution surface of the leached sediment. After this relatively rapid precipitation, the normal balance corresponding to the given stage of evaporite deposition would be restored, resulting in the deposition of salts alternating with anhydrite laminae. Repeated decreases in brine concentration would result in renewed leaching of chlorides from below , affecting the sediment surface most and losing intensity toward depth. Such cycles would provide sharp contrasts between the partially leached sediments and the overlying reprecipitated salt.

This interpretation of Herde (1953) seems to be supported by the existence of local sylvinite deposits overlying the eighth (swath) zone, as well as by the presence of the Riedel potash deposit directly overlying the uppermost swath zone. Both sylvinite zones indicate the existence of marginal depressions, in which the potassium salts leached from the underlying sediments were deposited after short transportation. Corroded to rounded sylvite crystals embedded in halite indicate that the transportation was in part mechanical.

Thus in these leached cycles the salts do not occur in successions of gradually increasing and decreasing solubility, as they do in the "normal" cycles, but highly soluble salts overlie hardly soluble and insoluble sediments with sharp contacts. The potash deposition does not mark the culmination of a continuously progressing evaporite deposition, but indicates the existence of local protected environments where potash salts could be deposited and preserved directly above deposits of the most recessive phase of the previous cycle.

As these cycles are the product of an interplay between terrestrial inflow and basin brines, they are not manifested uniformly throughout the entire basin, but are most distinct in a belt between the margin and the interior. The potash salts, repeatedly deposited and dissolved or reworked during these cycles, were occasionally redeposited in local depressions where they were protected from renewed reworking. Thus a proto-Riedel deposit may have been formed and destroyed several times during the deposition of the Swath Salt, as indeed indicated by the presence of corroded sylvite grains and even a local potassium salt deposit (Swath Deposit) over some of the swath zones (Herde, 1953), but only its last deposition was preserved over a larger protected area.

Figure 18

A swath cycle. Pure halite appears abruptly over the most contaminated top portion of the underlying cycle, and grades upwards into salt containing increasingly distinct dark laminae. The cross-bedding may be due to currents.
Hansa Mine, Hannover, Germany

The main feature of the depositional process was the permanent importance of terrestrial influences. In this extremely shallow brine, even minor terrestrial inflows invoked profound changes. From the start of the Swath Salt deposition, the brine frequently withdrew, laying dry extensive areas of the basin for short periods. A hexagonal network of gray anhydrite is common in the halite laminae of the swath zones and becomes increasingly distinct towards their top, indicating temporary subaerial desiccation. (This is not unusual in the Leine Cycle, where several instances of extremely shallow water and temporary desiccation are manifest both in the Main Anhydrite and in the overlying salt suite). The desiccational cracks of the swath zones often display a hexagonal pattern in vertical as well as in horizontal section (Fig. 18) and are filled mostly with gray anhydrite.

Frequently, the top of the cracks is filled with rounded sylvite crystals. Apparently, occasionally small shallow pools of highly concentrated brines flooded the suncracked halite; they dried out rapidly, depositing sylvite (Herde, 1953). The sylvinite was soon leached, unless it was washed into newly formed brine pools and open salt-cracks, where it remained preserved. Sometimes the sylvinite filling the top of the cracks continues upwards into a sylvinite layer, which, crystallizing from a brine pool, first filled the cracks in its bottom. In these cases, too, rounded sylvite grains frequently occur at the bottom of the sylvinite layer and in the crevices beneath, showing mechanical transportation from the surrounding dry areas.

Facial diversity was intensified by dissected bottom relief, perhaps the product of eroding currents. Below the network of desiccational cracks, characterizing the most contaminated top section of the swath zones, structures indicating paleocurrents can be observed (Personal observation, 1970). Anhydrite laminae, parallel to the layering, are connected (Fig.18) by inclined laminae of anhydrite,cutting across the layering and traceable downwards into the upper part of the pure salt zone. The inclined laminae are several meters long, parallel with each other, and meet the layering at an angle of 10-20^{o}. The slow currents responsible for their formation were apparently characteristic of the shallow coastal water prior to total emergence.

During the first cycle of the Swath Salt formation, most of the clay entering the basin was deposited in a marginal depression (Blue Shale); only minor clay flakes could pass its surrounding shoals and enter the main basin. During the following cycle of reworking and redeposition, potassiferous salts were collected in the same depression . (This marginal position of many potash deposits is emphasized by Stewart (1954). Except for this areal coincidence, however, the distribution of shoals and deeps was different for each cycle, though their differentiation increased. Presumably this increasingly dissected basin floor was responsible for the fact, that in contrast to the lower, Ronnenberg potash deposit of the Leine Cycle, which is continuous and of constant thickness, the Riedel deposit that overlies the uppermost swath zone is discontinuous, though it occurs over a larger area. Its constituting discontinuous potash deposits probably reflect the distribution of local deeps which trapped the heavy potassiferous brines.

The small amounts of Mg, even in the potash deposits, may indicate that direct marine influences were subordinate. (Herde, 1953). The halite was probably, at least in part, washed in from older halite beds, temporarily exposed around the shrinking basin.

Argillaceous salt, leached cycles

As indicated in the descriptive section, similar leached cycles characterized by gradually decreasing and abruptly increasing salt purity occur also in the argillaceous portions of the third and fourth Zechstein cycles. The shale forms irregular intercrystalline relics and networks as well as thin laminae in the salt. These shale structures increase upwards within each cycle both in size and frequency, and culminate in a shale bed abruptly overlain by the pure basal salt of the next cycle. The argillaceous beds are often characterized by mudballs. It is thought (Herde, 1953) that this gradual upward decrease and abrupt increase in salt purity is due to a period of leaching, perhaps by rain or terrestrial water acting upon desiccated salt or salt covered by only a thin sheet of brine, followed by a period of renewed precipitation. The shallow depths of the Zechstein basin, generally accepted for the upper Leine and Aller Cycles, agrees well with this interpretation. Most sedimentary features, like the presence of mudballs and of intercrystalline shale relics, admit similar interpretation for the argillaceous Salina salts in the axial region of the Appalachian Basin (Part II).

Kieseritic and polyhalitic halite

Kieseritic halite often takes the place of anhydritic halite directly below potash deposits. The kieserite gradually replaces anhydrite as the dominant sulphate mineral in successive sulphate laminae between the annual or undecennial halite beds. Near land areas, argillaceous laminae also intercalate; the alternation of clay, kieserite and halite indicates that these small cydes are controlled by periodic precipitation (Fig. 11).

More complex is the occurrence of polyhalite. Though it is frequently present as an alteration product of anhydrite, its only major occurrence is in the Stassfurt Cycle, in the Sub-Hercynian and Thuringian basins (Simon, 1970). The polyhalitic like the kieseritic halite grades from anhydritic halite. In the lower section of the polyhalitic suite, halite beds are separated by anhydrite and anhydritic polyhalite laminae; further up, these sulphate laminae consist essentially of polyhalite,and the halite itself contains dispersed polyhalite crystals; near the top, the sulphate laminae consist either dominantly of anhydrite or dominantly of polyhalite, and the latter also form thicker beds (1-20 cm). Similar characteristics of a polyhalitic Zechstein evaporite suite in England led Stewart (1963) to conclude that much of the polyhalite in the upper beds was deposited as a primary mineral, and that simultaneous metasomatism of anhydrite into polyhalite in the underlying sediments by downward percolating potassiferous brines provided most of the Ca that was needed for the primary polyhalite precipitation. Small cycles of

4. halite, with some polyhalite
3. polyhalite, with some halite
2. halite, with some anhydrite
1. anhydrite, partly replaced by halite

are interpreted by him as the results of probably seasonal changes in salinity and temperature. According to Stewart's tentative correlation (1954), (not followed by Lotze, (1957)), the ~~the~~ polyhalitic English and German evaporites are approximately contemporaneous.

From observations by Simon (1970) at Asse in the Sub-Hercynian basin it appears that the polyhalite formation, both primary and secondary, preceded the deposition of kieseritic halite, which, on its turn, introduced the Stassfurt potash deposit. It is significant that Stewart (1963), too, finds that the polyhalitic beds are overlain by kieseritic halite. At Fordon, England, however, where he studied the polyhalitic halite, the overlying potash deposit is absent, either due to non-deposition, or due to intraformational dissolution. The latter is suggested by the scarcity of kieserite near the top of the kieseritic zone, and the fact that this zone is overlain, with a sudden petrographic change, by anhydrite.

The intensive deposition of polyhalite (and the leaching) may reflect the marginal position of both the Sub-Hercynian and the Yorkshire basins and their proximity to dry lands. This is further emphasized by the local argillaceous facies overlying the polyhalite in the Sub-Hercynian basin (Asse Mine, Simon, 1970), deposited from muds swept in from the temporarily uplifted Eichsfeld shoal. The latter was probably connected to that flat island of the Harz region to which the red coloring and increased clay content of the similarly polyhalitic Stassfurt salt in the Mansfeld basin of Thuringia are attributed.

Sylvitic and carnallitic halite

The potash zones of the Zechstein have been the subject of numerous articles, therefore here we do not intend to deal with them in greater detail. The enormous potash deposit of the Stassfurt Cycle comprises the bulk of potassiferous rocks in the German Zechstein, occurring in lateral continuity over the largest area (100,000 km^2, Fig. 17) in the greatest thickness (50 m). It displays a distinct lateral zoning, with sylvinite and anhydrite along the basin margin, sylvinite and kieserite somewhat basinward, and carnallite-halite with kieseritic sylvinite (hartsalz) lenses in the basin interior. All areas are characterized by distinct cyclic sedimentation; the cycles consist of sylvinite and kieserite; halite, kieserite and carnallite; and halite and carnallite.

The lateral facies pattern is attributed to either primary or secondary causes. Lotze (1938) and Richter-Bernburg see in it primary facies distribution, and their approach is accepted by most geologists working in the German Zechstein. Borchert (1959,1963), however, assumes that carnallite had been the primary K-mineral throughout the basin, and attributes the zoning to late diagenetic leaching of Mg by waters escaping during dehydration from the basal sulphates underlying the Stassfurt salt. According to him, only the thick gypsum deposits of the basin margin could provide water in amounts sufficient to accomplish this leaching, since the sulphates of the interior are thin and were mostly deposited as anhydrite.

There are considerable objections to this theory, as discussed above (see: Flaser anhydrite genesis): there is no field evidence of any channel system that would have led the water from below the 500-m Stassfurt halite to the potash deposit above it; neither is there any dissolution along the contact of the $CaSO_4$ with the overlying halite.

The Ronnenberg potash deposit in the third Zechstein cycle consists largely of sylvinite; local cycles of alternating red and green sylvinite are observed. Hofrichter (1960) applied varve chronology to determine whether the potash deposit was time constant. With good correlation of sequentially numbered anhydrite laminae established in sections up to 20 km apart, he found that both the basal and the top contacts of the deposit occur at different laminae in different sections. Further study revealed that in certain regions, identified as deeper basin areas, the sylvite deposition started earlier and lasted longer. Thus lenticular sylvinite bodies were formed, whose top and bottom contacts cut across the time plane, but whose level of largest extension, corresponding to the time most favorable for potassium salt deposition, is time-constant. The differences in the starting times of sylvite precipitation at different localities were small: if the anhydrite laminae mark annual and not undecennial intervals, they rarely exceeded 10-20 years.

It is assumed (Hofrichter, 1960) that preferential sylvite deposition was controlled by its lower solubility in colder water. Accordingly, sylvite deposited longer in the deeper and colder parts of the basin, while the deposition of halite was less temperature-dependent and extended to the intervening shoals.

The original cyclic layering of the Ronnenberg deposit has been modified by the reworking of previously deposited sediments. Chemical and mechanical reworking is manifested by corroded, rounded crystals. Locally, even a "halite conglomerate" was formed: it consists of rounded halite crystals between which the sylvite has been leached. In the shoal areas, intervals of local desiccation are indicated.

The Riedel potash deposit consists of discontinuous lenses, supposedly deposited (Herde, 1953) in minor depressions collecting the heavy potassiferous brines. Gray and red sylvinite often interfinger with, and wedge out in, low-potassium sylvitic halite, the supposed sediment of the warmer shoal areas. Signs of repeated desiccation, partial and total dissolution, reworking, and redeposition in local brine pools are numerous; they have been briefly reviewed above in connection with the leached anhydritic halite (swath) cycles.

THE SILURIAN SALINA GROUP OF THE APPALACHIAN BASIN

Tectonic-palaeogeographic background:

Development of the Appalachian Basin prior to the Acadian Orogeny

The Appalachian Basin (Figures 1-2) is a northwest-trending sedimentary basin between the orogenic zone along the southeastern margin of North America to the southeast, and the Cincinnati- Findlay-Algonquin arch system merging into the Canadian Shield to the northwest. Prior to the formation of the Taconic Mountains in the Ordovician, the basin had sloped straight into the ocean, while the arch system was also less distinct. Even later, some communication did exist between the Appalachian and the neighbouring Michigan Basin through a gap (Chatham Sag, Alling and Briggs, 1961) between the Findlay and Algonquin arches, and frequently also above the arches.

The basement was most mobile close to the geosyncline and subsequent mountain system, and became increasingly cratonic toward the arch system. It was thus a typical foredeep, grading from the platform into the geosyncline before the orogeny and later separated from the ocean basin by the newly formed mountain system.

Within the basin, three tectonical zones can be distinguished. The northwestern, platform zone is underlain by a tectonically in-active basement, and forms a flat monocline whose layers are dip-ping to the southeast. The Palaeozoic sequence is relatively

<u>Fig. 1</u>

Tectonic map of south-eastern North America during the Cambrian, Ordovician and Silurian (after Eardley, 1951). 1-Appalachian Basin, axial zone; 2- Michigan Basin; 3-Generally emergent, eroding areas; 4-Stable interior, receiving less than 5000 ft of sediments

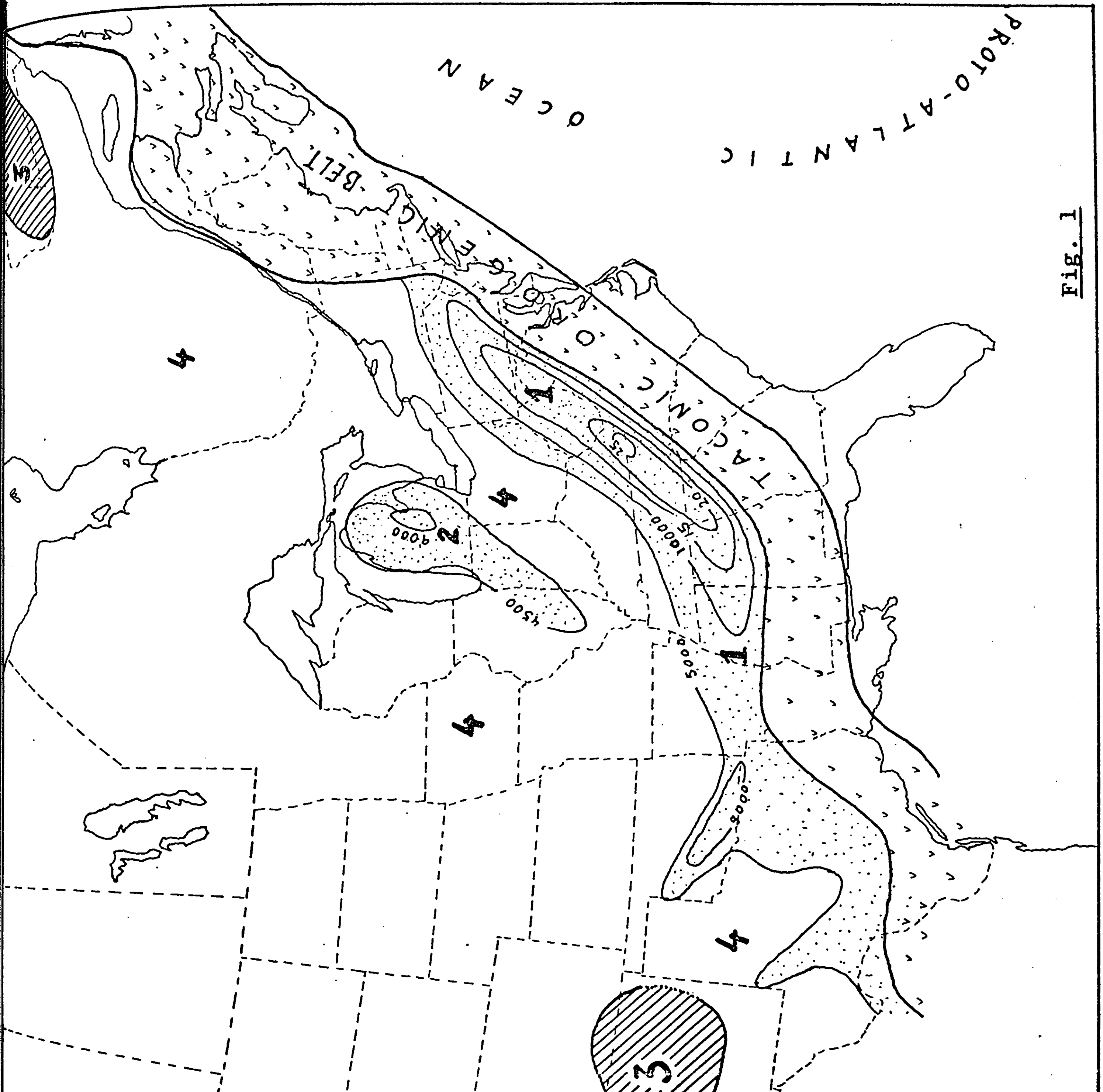

PROTO-ATLANTIC OCEAN
TACONIC OROGENIC BELT
1
2
3
4
25
20
15
10000
5000
9000
4500
5000

Fig. 1

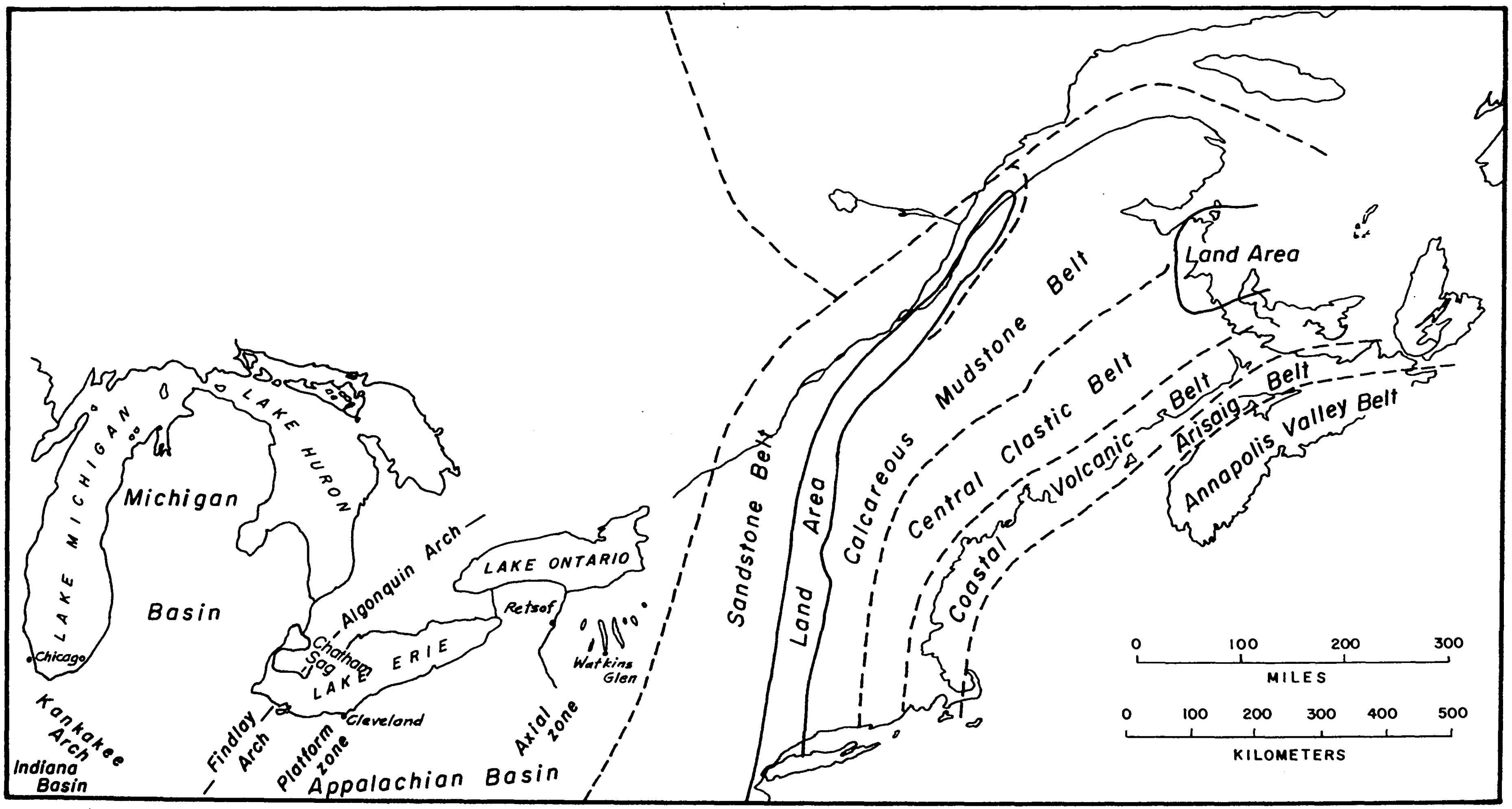


Fig. 2. Relationship between the evaporite-depositing Michigan-Appalachian basin complex in the west and the Taconic orogenic belt in the east during Upper Silurian Ludlow times. The belt of coarse clastics surrounding the mountain-islands grades to the west into evaporite-bearing sequences. After Boucot, 1968, and Rickard, 1969

thin, most so along the arch system where post-Ordovician sediments are largely absent. The tectonic regime is similar to that on the western slope of the arch system, resulting in sedimentary features similar to those observed in the marginal areas of the Michigan Basin.

To the southwest of the extensive platform zone is a sedimentary trough, 20-40 miles wide, containing a thick and only slightly deformed Palaeozoic sequence; this is the present axial zone.

Past axial zones were, however, somewhat further to the southeast, in the orogenic zone, where two subzones can be distinguished: the miogeocline belt of the pre-Taconic monoclinal shelf sequence (called so by Bird and Dewey because it did not form a separate syncline), and the adjacent eugeosynclinal belt characterized by extensive volcanism. Maximum thicknesses of the Lower Palaeozoic sequence are over 22.000 ft close to the boundary between the orogenic and the present axial zones.

The distribution of sedimentary facies and thicknesses reflects this tectonic pattern. Prior to the Taconic Orogeny, in the Cambrian, the detritus was supplied to the basin mainly from the loose mantle previously accumulated over the Precambrian of the Canadian Shield, forming a sandstone belt along the northern margin of the Michigan-Appalachian basin complex. During the late Cambrian and the Ordovician, however, the flat shield became increasingly inundated, while the mountainous islands of the coastal zone increasingly emerged, leading to modifications in the facies pattern. In the orogenic zone, sand was deposited, partly in a marine, and partly in a continental environment. The axial zone of the basin was generally characterized by shale and carbonate sedimentation, the shale decreasing and the carbonate increasing with increasing distance from the orogenic zone. Finally, in the platform areas, mainly dolomites were formed.

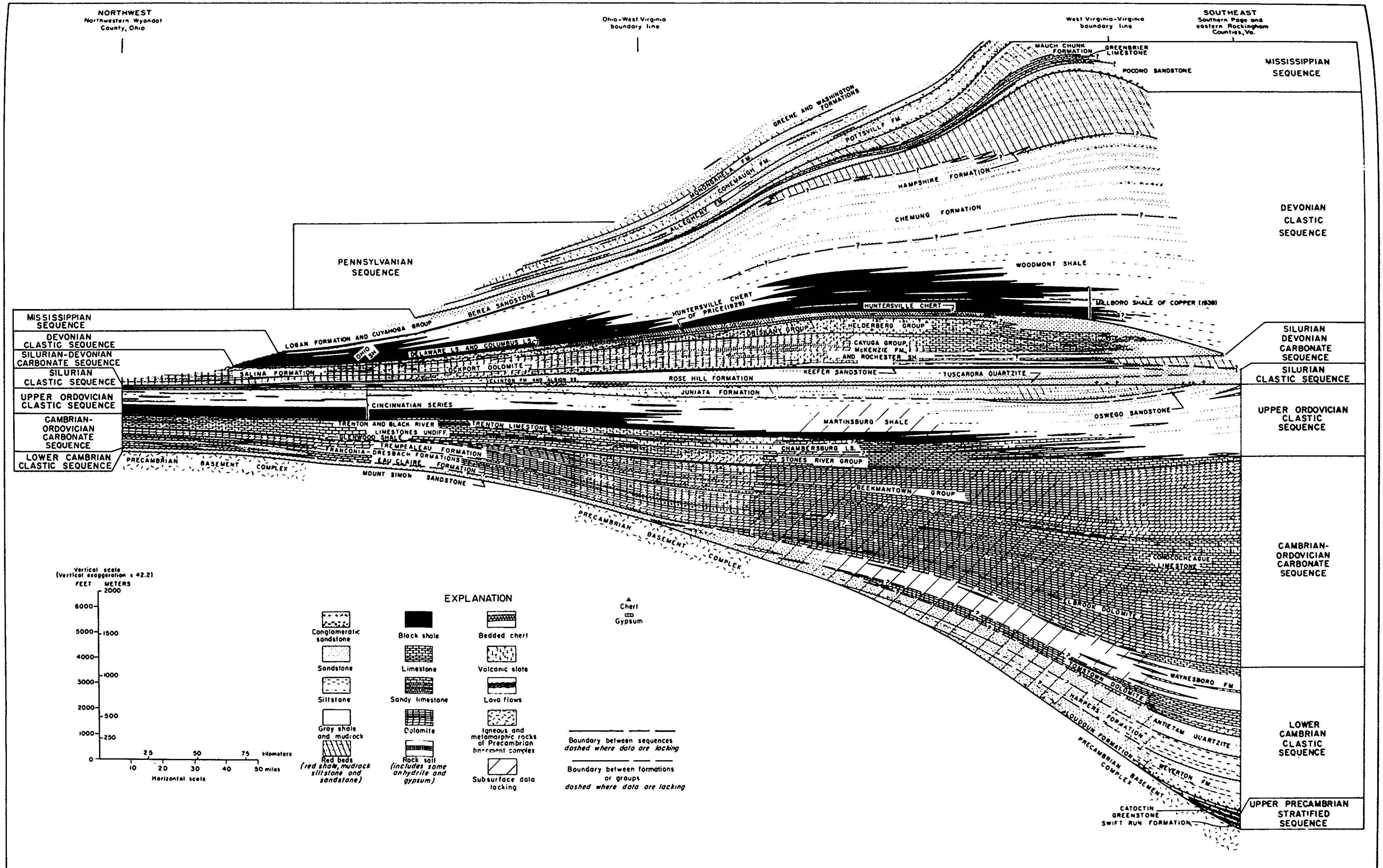
Precambrian Basement

Information about the Precambrian basement of the basin is still sparse, so most evidence is indirect, obtained from the bordering areas of the Canadian Shield. This extrapolation, however, is justified, as these areas had originally formed part of the basement of the Appalachian Basin, and were only exposed by late erosion of the Lower Palaeozoic sequence.

The earliest radiometric ages known from the infrastructure which includes the Appalachians are Pb/U ages of 1145 and 975 m.y. from the Bear Mtn. in New York. (Lyons and Faul, 1968). These ages were recorded in granites intruding into the Grenville series, a shelf sequence 10.000 ft thick, now consisting chiefly of metamorphosed limestones and cross-bedded sandstones. Tillites overlying this series were formed close to the end of the Precambrian (about 600 m.y. ago).

Fig. 3.

Stratigraphic relations of the Appalachian basin fill (from Colton, 1970). Note rapid decrease in thickness toward the platform zone in the west, predominance of carbonates before and clastics after the Taconian orogeny in the Ordovician, and intercalation of the Salina evaporite suite between the upper Ordovician-Silurian (syn- and post-Taconian) and Devonian (syn- and post-Acadian) clastic sequences.


NORTHWEST
Northwestern Wyandot County, Ohio
Ohio-West Virginia boundary line
West Virginia-Virginia boundary line
SOUTHEAST
Southern Page and eastern Rockingham Counties, Va.
MISSISSIPPIAN SEQUENCE
DEVONIAN CLASTIC SEQUENCE
SILURIAN-DEVONIAN CARBONATE SEQUENCE
SILURIAN CLASTIC SEQUENCE
UPPER ORDOVICIAN CLASTIC SEQUENCE
CAMBRIAN-ORDOVICIAN CARBONATE SEQUENCE
LOWER CAMBRIAN CLASTIC SEQUENCE
PENNSYLVANIAN SEQUENCE
SILURIAN DEVONIAN CARBONATE SEQUENCE
UPPER PRECAMBRIAN STRATIFIED SEQUENCE
MAUCH CHUNK FORMATION
GREENBRIER LIMESTONE
POCONO SANDSTONE
GREENE AND WASHINGTON FORMATIONS
MONONGAHELA FM.
CONEMAUGH FM.
ALLEGHENY FM.
POTTSVILLE FM.
HAMPSHIRE FORMATION
CHEMUNG FORMATION
WOODMONT SHALE
MILLBORO SHALE OF COOPER (1939)
BEREA SANDSTONE
HUNTERSVILLE CHERT OF PRICE (1929)
HUNTERSVILLE CHERT
LOGAN FORMATION AND CUYAHOGA GROUP
OHIO SH.
DELAWARE LS. AND COLUMBUS LS.
ORISKANY GROUP
HELDERBERG GROUP
CAYUGA GROUP, McKENZIE FM. AND ROCHESTER SH.
LOCKPORT DOLOMITE
SALINA FORMATION
CLINTON FM. AND ALBION SS.
KEEFER SANDSTONE
TUSCARORA QUARTZITE
ROSE HILL FORMATION
JUNIATA FORMATION
CINCINNATIAN SERIES
OSWEGO SANDSTONE
MARTINSBURG SHALE
TRENTON AND BLACK RIVER LIMESTONES UNDIFF.
TRENTON LIMESTONE
GLENWOOD SHALE
TREMPEALEAU FORMATION
FRANCONIA-DRESBACH FORMATIONS
EAU CLAIRE FORMATION
CHAMBERSBURG LS.
STONES RIVER GROUP
MOUNT SIMON SANDSTONE
PRECAMBRIAN BASEMENT COMPLEX
BEEKMANTOWN GROUP
CONOCOCHEAGUE LIMESTONE
ELBROOK DOLOMITE
WAYNESBORO FM.
TOMSTOWN DOLOMITE
ANTIETAM QUARTZITE
HARPERS FORMATION
LOUDOUN FORMATION
WEVERTON FM.
CATOCTIN GREENSTONE
SWIFT RUN FORMATION
Vertical scale (Vertical exaggeration x 42.2)
FEET
METERS
6000
5000
4000
3000
2000
1000
0
2000
1500
1000
500
250
25
50
75
kilometers
10
20
30
40
50 miles
Horizontal scale
EXPLANATION
Conglomeratic sandstone
Sandstone
Siltstone
Gray shale and mudrock
Red beds (red shale, mudrock siltstone and sandstone)
Black shale
Limestone
Sandy limestone
Dolomite
Rock salt (includes some anhydrite and gypsum)
Bedded chert
Volcanic slate
Lava flows
Igneous and metamorphic rocks of Precambrian basement complex
Subsurface data lacking
Chert
Gypsum
Boundary between sequences dashed where data are lacking
Boundary between formations or groups dashed where data are lacking

At the end of the Precambrian, the Grenville Orogeny lead to the formation of the Appalachian geosyncline and of the related proto- Atlantic ocean. Radiometric data for this event, obtained from many Precambrian exposures along the Appalachians, give unexpectedly young ages ranging from 468 to 640 m.y. and averaging at about 540 m.y. (ibidem).

According to Bird and Dewey (1970), this orogeny resulted in the formation of a rift valley dividing the continent of the northern hemisphere into a Canadian and a Eurasian segment. The two segments moved apart on divided lithosphere plates, while the new ocean floor between them was spreading by accretion along a mid-oceanic ridge system as in the present Atlantic ocean. The coastlines of the resulting proto-Atlantic ocean displayed tectonic features like those of the present Atlantic coast of the American continent, with the eroded Precambrian basement becoming increasingly buried under thick sedimentary sequences deposited over the lower Palaeozoic continental shelf (miogeocline) and slope (eugeosyncline).

Basin fill (Fig.3)

Cambrian

The Cambrian sequence of the Appalachian Basin is about 8000 ft thick at the northwestern margin of the orogenic zone , and decreases toward the Canadian Shield. Additional basin centres, with thicknesses of 2000-3000 ft, occur in Michigan and Indiana. Prolonged previous weathering over the Canadian Shield had produced large amounts of accumulated detritus , which supplied the material for a belt of sandstones, 100 miles wide, along the northern margin of the Appalachian and Michigan basins, deposited mainly during the early Cambrian. The erosion of highlands formed by the Grenville Orogeny along the southeastern boundary of the Appalachian Basin also provided sediments, but in smaller amounts. Generally, the central and southern part of the Appalachian Basin, south of the sandstone belt, was characterized by the deposition of carbonates with shale interbeds, while in the northern part of the Appalachian Basin and in the Michigan Basin sandstones play a significant role. A basal sheet of sandstones, metamorphosed into orthoquartzites, can be traced throughout the area of Cambrian sedimentation.

Along the continental margin, Theokritoff (1968) distinguished between a western shelf and an eastern basin sequence. The former is characterized by sandstones and sandy carbonates, transgressively overlying Precambrian massifs. The basin sequence of the orogenic zone, to the east, reflects the continental slope and rise of the proto-Atlantic ocean and consists now essentially of phyllites, schists, gneisses, graywackes, and greenstones. The presence of non-volcanic islands could be demonstrated, otherwise the series was found to increase in thickness toward the contemporary ocean in the southeast.

By the middle Cambrian, the amount of detritus decreased throughout the Appalachian Basin. The late Cambrian was characterized by extensive transgression beyond previous coastlines, though total submergence was probably not reached until the middle Ordovician for the entire North American platform.

Ordovician

While during the late Cambrian or early Ordovician a minor orogeny (Penobscot; Neuman, 1967) in Maine and possibly in Quebec may have been a precursor to the Taconic movements, in other parts of the geosyncline sedimentation continued unaffected. Osborne and Berry (1966) have shown that the deposition of the Levis shales in Quebec, which can be followed to New York (Berry, 1968), was continuous from late Cambrian to middle Ordovician times.

Toward the platform, this shale belt is bordered by a belt of limestones and dolomites, with frequent mudcracks, desiccation breccias and Cryptozoon algae, covering part of the Appalachian Basin, while to the east it grades into a coastal belt of islands and volcanoes, a belt of sandstones and conglomerates with tuffs and lava flows. Carbonate deposition in the limestone-dolomite belt remained continuous during the early Middle Ordovician in probably grabenlike small basins, while the interposed horsts were subjected to submarine erosion (Zen, 1967).

Movements of the Taconic Orogeny had already started during early middle Ordovician times. Folding, uplift and erosion took place over much of New England, New York, and adjacent Canada, where late middle Ordovician formations unconformably overlie lower Ordovician sediments and frequently start with a basal conglomerate. The red shales and cherts at the base of the Normanskill formation in New York may represent fossil soil that developed on such a land (Berry, 1968).

By the Middle Ordovician the belt of coarse clastic sediments, bordering the Michigan-Appalachian basin complex from the north during the Cambrian, had withdrawn to the west of the Lake Michigan, while to the north it was absent to the erosional boundary with the Canadian Shield, indicating almost complete inundation of the continent. Both basins became filled with carbonates, over 600 ft thick along the western margin of the orogenic zone and more than 500 ft in the centre of the Michigan Basin.

During late Middle Ordovician times the area of New England and neighbouring parts of Canada became relatively mobile with platforms and ridges separating relatively deep troughs. Lands uplifted during early Middle Ordovician times became widely submerged. At the same time a larger land area became uplifted in western Massachusetts and southern Vermont between the shale belt and the belt of volcanoes and islands, while a larger land mass, probably a long coastal island, existed in eastern Maine and in New Brunswick. These two land zones show angular unconformities which grade through disconformities into conformable sequences both toward the ocean basin on the eastern and the Appalachian Basin on the western side of the orogenic zone. In addition, a narrow trough of conformable calcareous flysch deposition occurred between the two land areas.

In its type area, the Taconic allochthon, the Taconian orogeny occurred during the late Middle Ordovician Trenton time (Caradocian in the European time scale) and involved two major events: An initial event of normal longitudinal faulting produced a submarine horst and graben topography in the eugeosynclinal sequence, and a subsequent gravity sliding to the east resulted in the emplacement of an eugeosynclinal allochthon in the miogeoclinal shelf sequence of the late Middle Ordovician Normanskill sea. (Pavlides et al., 1968).

Due to these extensive uplifts, coarse-grained graywackes were formed in part of the shale belt, while muds of the Utica shales deposited over carbonates over most of the Appalachian Basin. Toward the west, these shales first become thinner, then grade into limestones along the western margin of the Appalachian Basin and into dolomites toward the Mississippi Valley, reflecting the general lithofacies pattern of the lower Palaeozoic.

Unconformities to the south, in Pennsylvania, indicate somewhat later, late Ordovician uplift and folding during late Maysfield (uppermost Caradocian) to Richmond (Ashgillian) time (Pavlides et al., 1968)

The same land area in western Massachusetts and southern Vermont, which provided the source of the late Middle Ordovician coarse clastics, may have been the source terrain for at least some of the Late Ordovician conglomerates and sandstones that comprise the Queenston delta, a precursor of the Bloomsburg delta which was simultaneous with part of the Upper Silurian evaporite deposition. These coarse clastics were deposited in the Appalachian Basin during early Late Ordovician times, but as erosion increased in the rising orogenic land areas, the shoreline moved westward until a low delta plain stretched from the foothills of the Taconian Mts. to beyond the Niagara, filling most of the Appalachian Basin. The Oswego conglomerates, Juniata sandstones and red Queenstone shales of this delta complex pass toward the west into the dark marine Richmond shales which replaced the limestones deposited during the Middle Ordovician.

These Upper Ordovician shales, with sandstones near the orogenic zone, covered almost the entire area of the Michigan and Appalachian Basins and graded into carbonates along the southwestern, western and northern margins of the basin complex. To the northwest of the Lake Michigan, a partly eroded narrow shale belt may indicate that some detritus was again supplied by emerging areas of the Canadian Shield.

Along the western margin of the orogenic zone, the total thickness of the Upper Ordovician shales and sandstones is more than 6000 ft; it decreases to less than 1000 ft toward its contact with the northern limestone belt, and to less than 500 ft in the shales covering the southern margin of the Canadian Shield.

The total Ordovician sequence, of which the detritic Upper Ordovician forms the greatest portion, ranges in thickness from more than 7000 ft along the orogenic zone to over 1500 ft in the centre of the Michigan Basin, wedging out toward the eroded areas of the Canadian Shield. Shales are the dominant sediments for most of the Appalachian Basin and carbonates with subordinate shales for the Michigan Basin, both grading into purer carbonates toward the central regions of the North American platform.

According to Bird and Dewey (1970), the Taconic Orogeny marked an important change in the movement of the lithospheric plates: the spreading of the proto-Atlantic ocean, continuing since the late Precambrian, stopped, and the lithospheric plates carrying the North American and European continents started moving toward each other. This change in plate movement transformed the eastern coast of the North American continent from Atlantic into Pacific type, with corresponding trough of plate consumption and intensive tectonism and volcanism. The process of ocean floor consumption lasted, according to these authors, throughout the Silurian, until in the Middle Devonian the Acadian Orogeny signalled continental collision.

Silurian

During the Silurian North America consisted of a broad platform, bordered to the east by the Appalachian geosyncline, to the north by the Franklinian geosyncline, to the west (except in California) by the Cordilleran geosyncline, and to the south by the Ouachita geosyncline. The bulk of the platform, submerged since the Middle Ordovician, was characterized by dolomites, encircled by a belt of limestones and calcareous shales and bordered by the geosynclinal rim. (Boucot et al., 1968).

During the early and middle Silurian times the Appalachian Basin received abundant sandy sediments from the east as the highlands of the Taconic ranges were gradually worn down. This material, trapped between Appalachia and the Cincinnati Arch, kept the whole basin silted up near to or above sea level, oscillating between shallow sea floor and coastal plain.(Dunbar, 1960).

The narrow coastal islands of Appalachia were bordered toward the Appalachian Basin by a belt of non-marine sands, 50-100 miles wide, now resembling the Old Red Sandstone of Great Britain. This sandstone belt moved, with advancing transgression, gradually to the east, finally covering the land belts themselves during late Llandovery and Wenlock times.

To the west the sandstone belt graded into a belt of calcareous mudstones, traceable southward to Alabama and containing in the Middle Silurian Clinton time a lenticular bed of oolitic haematite, generally 1 inch to 4 ft, in Alabama 40 ft thick. These iron accumulations, like the evaporites of the late Silurian, are interpreted as deposits of basins with increasingly inhibited circulation, bordering an increasingly peneplenized mountain system with waning detritus supply (Hunter, 1970).

The western, platform zone of the Appalachian Basin and the whole Michigan Basin was, particulary during the Middle Silurian, characterized by carbonate sedimentation. Limestones graded into dolomite both upward in the sedimentary column and toward the central areas of the North American platform. Growing on calcarenite mounds, flat Bryozoan reefs were overgrown by over 70 ft high reefs of Tabulata, Tetracoralla and Stromatoporoidea.

Contact with the open sea, already restricted by the assumed advanced contraction of the proto-Atlantic ocean, by the Taconic orogeny, and by the ensuing accumulation of detritic sediments, became in Michigan further obstructed by the appearance of high reefs in the Middle Silurian. To this a minor uplift may have contributed, elevating the reef tops above sea level and stopping their further growth (Gill, oral comm., 1970). In the Appalachian Basin major reef growth was inhibited by the

the larger inflow of clastics into the basin; here the development of the Bloomsburg delta during early Late Silurian times may have been the major factor in restricting circulation.

Due to these factors and the ensuing arid climate, the areas of carbonate and alternating carbonate-mudstone deposition in the Michigan and Appalachian Basins became, by the start of the Late Silurian (late Ludlow), the site of evaporite deposition. In a sequence of evaporites alternating with dolomite and shale beds, evaporites (chiefly halite) of over 2000 ft and 800 ft, respectively, were deposited in the Michigan and northern Appalachian Basins,while the southern and eastern areas of the Appalachian Basin were characterized by continued limestone-mudstone deposition. The combined area of the tow basins characterized by evaporite deposition is often designated as the Salina Basin for this period.

During late Pridoli times the evaporite sedimentation ceased, first in the Appalachian, then also in the Michigan Basin, and argillaceous dolomites and coralliferous limestones formed until the end of the Silurian. In New York, the Siluro-Devonian boundary is within the Rondout dolomite formation (Berdan et al., 1969).

In the geosynclinal region, to the east of the Taconic uplifts, the facies pattern is more complex. According to Boucot (1968), a narrow belt of marine clastics bordered the orogenic island of Appalachia from the east. Along the axis of this belt, calcareous flysch deposited during the Taconic Orogeny, grading from late Llandovery into the sand and silt characteristic of the entire clastic belt. The the east of the flysch belt, a zone of graywackes and argillaceous sandstones is the locus of the thickest Silurian sequence in the northern Appalachian region.

In early Llandovery times, this belt of coarse clastics was bordered to the east by a narrow land-strip, which later submerged to be replaced by an area of marine sedimentation and intensive volcanic activity throughout the remainder of Silurian time.

The total thickness of Silurian sediments is over 5000 ft at the western margin of the orogenic zone and over 3500 ft in the Michigan Basin, but decreases to less than 750 ft over the arch system separating the Michigan and Appalachian Basins.

Early Devonian

Two distinct periods can be differentiated in the Devonian sedimentation of the Appalachian region: the period leading up to the Acadian Orogeny and the subsequent period of syn- and post-orogenic sedimentation. Only the first period will be considered here.

The Lower Devonian consists mainly of carbonate rocks,with subordinate sandstones: the Manlius, Coeymans and New Scotland (lower and middle Helderberg) formations comprise the Gedinnian, the Becraft (upper Helderberg) and Oriskany formations the Siegenian, and the Esopus and Schoharie formations the lower Emsian stages in the European time scale.

In the Appalachian region,(Boucot, 1968) the middle and late Helderberg are characterized by the emergence of a major land area where the marine Arisaig belt and the coastal volcanic belt had existed from late Llandovery to early Gedinnian times. The northeastern part of this land area is characterized by volcanic rocks; the southeastern part, in Nova Scotia, contains red beds only. The major island of Appalachia reached to the north only to northern New Jersey and southeastern New York; it supplied quartz sand to the Appalachian Basin. North of this land area, limestone of the platform type occurs to the west, whereas a calcareous siltstone and limestone belt occurs to the east and grades southeasterly into a belt of shales and siltstones, containing a region with volcanic deposits.

During Oriskany and Schoharie times platform-type carbonates and sands grade to the southeast through a narrow belt of fossiliferous shallow-water marine calcareous sediments into widespread cyclically banded clays and sands, bordered to the southeast by a non-marine environment apparently characterized by volcanism. The platform-type quartzose sands form a wide belt about 100 miles across, bordering the temporarily emerged Canadian Shield. (Boucot, 1968).

Towards the Appalachian Basin, the Helderberg rocks are generally confined to the eastern zones and occur mainly in carbonate facies. Quartzose sandstones of the Oriskany extend to the whole basin, as does also the lower Middle Devonian Onondaga formation which frequently contains reefs. Similar coralliferous carbonate rocks occur also in the Michigan Basin, replaced toward the basin centre by a lower Middle Devonian evaporite-bearing series, 600 ft thick, which contains evaporite beds of more than 300 ft aggregate thickness. (Snyder, 1955).

The discovery of marine Middle Devonian (Eifel) rocks, mainly limestones and shales, in the northern Appalachians (Boucot and Johnson, 1967) demonstrated that the Acadian orogeny started at the boundary of the Eifel and Givetian and not in the early Devonian as previously supposed. This orogeny was, according to Bird and Dewey (1970), due to continental collision, marking the end of plate consumption in the proto-Atlantic ocean. In terms of its areal extent, the depth of the crustal units involved, and the intensity of metamorphism, it was (Zen, 1968) the most important diastrophic event in the northern Appalachian region. It was the time of at least one major episode of formation of nappes and of evolution of synclinoria and anticlinoria, and a period of emplacement of igneous rocks, both intrusive and extrusive; these igneous activities persisted from preorogenic to postorogenic times.

Tectonic facies

In concluding this brief discussion it may be noted that the Palaeozoic sedimentation in the Appalachian Basin and neighbouring regions took place in two major tectonic facies: Platform carbonates with few, mostly basal, quartzose sandstones, were deposited from the late Precambrian to the middle Ordovician Taconic Orogeny throughout the entire area between the gradually submerging Canadian Shield and the continental-geosynclinal margin. These carbonates became deformed along their eastern margin during the Taconic and subsequent orogenies, after which their eastern limit of deposition gradually receded to the west to give place to detritic sediments coming from the newly formed mountain ranges.

Foreland molasse, generally wedge-shaped bodies of red-beds grading into dark shales toward the interior of the basin, formed from the start of the Taconic Orogeny in the Middle Ordovician. After the Acadian Orogeny in the middle and late Devonian they became dominant throughout the basin.

The Silurian evaporites were formed between the Taconic and Acadian orogenies among platform carbonates and along the boundary of the platform and molasse sequences. Their formation was followed by a short lasting transgression of the platform carbonate suite to the east (late Silurian - early Devonian) due to the final erosion and submergence of the Taconic mountain ranges. After the Acadian Orogeny, the centres of evaporite formation receded with the platform carbonates westwards to the Michigan and Williston Basins in the Devonian and to the Williston Basin in the Mississippian.

The Upper Silurian Salina Group

The Upper Silurian (Cayugan) of the northeastern United States can be subdivided into two major units. The lower one, which consists of the Salina Group and its time-equivalents, is characterized by widespread presence of evaporites and clastic rocks (mainly shales), while in the upper unit shallow water carbonates are predominant.

The sediments of the Salina Group and its equivalents generally overlie shelf carbonates (Upper Niagaran Guelph-Lockport and McKenzie formations); dolomites and limestones with frequent bioherms and biostromes. To the east, close to the orogenic zone, these carbonates grade into calcareous shales, and finally, at the eastern basin margin, become undistinguishable from the red non-marine shales and sandstones of the lower Salina Group. (The term Salina is consistently used here in the sense of a time unit, including time-equivalents of the Salina Group of New York.)

The coarsest detritic sediments (sandstones, conglomerates, and red silty shales) occur in the east, along the western Piedmont of the eroding Taconic Mountains, in New Jersey, eastern New York and eastern Pennsylvania. They are part of the continental molasse suite, overlying partially similar clastic sediments ranging in age from middle Ordovician to early Silurian; their fauna is restricted to rare single beds and consists of ostracods, Lingula, and fish.

Toward the northwest and west, in the axial zone of the Appalachian Basin (central New York and Pennsylvania), and with advancing time and transgression in the east as well, the non-marine molasse grades into a sequence of variegated and dark shales with dolomite and evaporite interbeds. The fauna consists mainly of arthropods and brachiopods, the frequent eurypterids often reaching 10 ft in length.Trilobites, cephalopods, and graptolites, presumably washed in from the open sea, are rare, and so is sthenohaline benthos (reef-builders like stromatoporoidea with the less sensitive algae). In the vicinity of evaporites, the fauna is further impoverished.

Further to the northwest and west, in the platform regions, of Ohio and Michigan, shales become gradually subordinated to the dominant dolomite-evaporite sequence. The fauna is generally the same as in the variegated and dark shale sequences (Allings and Briggs, 1961), except that eurypterids occur less often. Here, too, in the vicinity of evaporites the fauna further decreases.

This belt, commencing with shales in the east and grading into dolomites to the west, is bordered from the south by a similar belt in which limestones are dominant. Here, too, there is a decrease in clastics from New Jersey, southern Pennsylvania, and Virginia, where sandstones and red shales are significant, to West Virginia and Indiana, where limestones are exclusive. In eastern Tennessee, a sandstone belt occurs within the limestone sequence. As in the north, the sandstones and red shale normally contain no fauna, while the limestones and calcareous shales (Fig.2) display abundant benthos, consisting of articulate brachiopods, eurypterids, stromatoporoids and corals.These latter forms occur only rarely in the north until they return at the end of Salina time, marking the end of evaporite deposition and the return to normal marine conditions. It may be significant that this return coincides with a marked decrease in detritus supply, due to the submergence of the major remnants of the mountain-islands formed during the Taconic orogeny.

Subdivision of the Salina Group

Since most evaporites are leached from surface exposures, subdivision of the evaporite-bearing sequence had to be based on subsurface observations, providing necessarily a system different from the one based on surface outcrops. At the same time it had to be based on lithostratigraphic criteria alone, because the poor fauna of the sequence is of little stratigraphic value.

The cyclic nature of the evaporite-bearing sequence, consisting of alternating argillaceous carbonate and halite suites, provided the basis for the lithostratigraphy established by Landes (1945). Correlating wells throughout Michigan, he differentiated the following major units:

Unit H (Bass Island dolomite): 175-570 ft thick, largely buff dolomite, some gray dolomite near base, anhydrite and salt beds near center of basin

Unit G: Uppermost Salina formation, 4-100 ft thick, characteristically gray shaly dolomite, green and red shales near Mackinac Straits

Unit F: Uppermost salt in the Salina section, 0-1230 ft thick. Thick beds of salt, separated by shale, shaly dolomite, and anhydrite

Unit E: 30-122 ft of gray or red shale with some dolomite, shaly dolomite and anhydrite

Unit D: 25-65 ft, nearly pure salt, with thin partings of buff dolomite

Unit C: 60-160 ft, largely shale or shaly dolomite with anhydrite and buff dolomite in places

Unit B: 240-260 ft thick, almost pure salt with minor dolomite

Unit A: 30-1105 ft thick, limestone, dolomite, salt, and anhydrite. This unit was later subdivided by Evans (1950):

A_2: Buff to brown dolomite or limestone
Dark gray dolomite
Salt or anhydrite

A_1: Fine brown to grayish brown dolomite or limestone
Fine gray dolomite or limestone
Salt or anhydrite

This subdivision has been further developed and partially modified by further studies, while their range of applicability became substantially extended. Significant tectonic aspects of the correlation have been discussed in several papers by Jacoby. Presentation of the distribution of Silurian salt deposits in New York by Kreidler (1957) and in Pennsylvania by Fettke (1955) as well as earlier petrographic studies by Alling (1928) for New York and more recent ones by Dellwig (1955) for Michigan have laid the foundations for the first regional syntheses: Landes (1960) has shown the distribution of salt beds throughout the area of deposition both in the Michigan and Appalachian Basins, and Alling and Briggs (1961) provided the first stratigraphic-palaeogeographic evaluation for the whole area. Drilling during the sixties resulted in the accumulation of new data. Stratigraphic evaluation of these by Cate (1961,1963, 1965) and Fergusson and Prather (1968) in Pennsylvania, Ulteig (1964) in Ohio, Hewitt (1962), Pearson (1963) and Sanford (1965) in Ontario, and Ells (1962,1967) in Michigan, as well as his own studies in New York, enabled Rickard (1969) to apply the subdivision of Landes, somewhat modified, to the correlation of widespread subsurface and some surface exposures throughout the Appalachian Basin. His study, encompassing more than 400 wells, is centred on the detailed correlation of gamma-ray logs, and attributes formation value to groups of subsurface lithostratigraphic units, correlating them with subdivisions formerly established on the basis of surface outcrops. Thus he obtains the following stratigraphic categories:

- Salina Group
  - Bertie Formation
  - Camillus Formation
    - Unit G (mainly shale)
  - Syracuse Formation
    - Unit F (mainly halite)
    - Unit E (mainly dolomite and shale)
    - Unit D (mainly halite and shale)
  - Vernon Formation
    - Unit C (mainly shale)
    - Unit B (mainly shale and dolomite)
    - Unit A (mainly shale)

The following analysis of macroscopic and microscopic sedimentary structures and cyclic sequences, based primarily on continuous cores penetrating the entire evaporite-bearing sequence at Watkins Glen, New York, in the axial zone, as well as on mines at Retsof, New York, in the axial and at Cleveland, Ohio, in the platform zone of the Appalachian Basin (all operations of International Salt Co.), will be presented in this regional framework.

Vernon Formation

In the continental molasse of the Taconic piedmont in the east, the lower portion of the Salina Group is represented by the Bloomsburg delta (eastern Pennsylvania), which consists of undifferentiated red siltstones with sandstone interbeds. To the northwest this sequence, up to 2000 ft thick, grades into a thinner sequence of gray and green shales, dolomites and evaporites, the Vernon formation. The reference section in Oneida Co., New York, consists of three units. An estimated 190 ft sequence of red shales, the lower unit, is overlain by gray and green shales and dolomites, containing a few beds of red shale and green sandstone, and, at top, a small marine fauna of brachiopods, gastropods, pelecipods, cephalopods, ostracods, fish, and eurypterids. This central interval, 89 ft thick, is overlain by about 140 ft of red and green shales, the upper unit. These three units, exposed in surface outcrops, were found by Rickard (1969) to be equivalents of the three lowest units (A, B and C) of the Salina Group, which are, in changing lithofacies, traceable to the Michigan Basin.

Lower Vernon (Unit A)

The base of the Vernon Formation is marked by the appearance of shale and anhydrite above the coarsely crystalline dolomites of the upper Middle Silurian. The top 60-70 ft of the latter, however, is also replaced by shales with interbedded dolomites toward the Taconic Piedmont in eastern New York and eastern Pennsylvania. Of the undifferentiated red shales and siltstones of the Bloomsburg delta, Unit A equivalents may comprise 700 ft(Rickard, 1969); this thickness gradually decreases to the west as the red shales grade into interbedded grey or green shales and dolomites. A broad platform along the Ohio-Pennsylvania state line , with less than 100 ft of sediments, separated a small part of the Appalachian Basin in northern Ohio, where a thinner sequence of carbonates and shales was deposited. Widespread deposition of evaporites, characteristic of Unit A in the Michigan Basin, was prevented in the Appalachian Basin by the influx of terrigenous sediments.

Middle Vernon (Unit B)

In the undivided red shales along the eastern margin of the basin, Unit B equivalents probably exceed 300 ft (Rickard, 1969). To the west, they grade into gray or green shales with dolomite interbeds, which contain the marine fauna mentioned above. Further to the west, dolomites become predominant. In a belt 50-100 miles wide, parallelling the Appalachian trend in the east and turning to the northwest in the west in northern Ohio, up to seven salt beds occur; their aggregate thickness is about 75 ft in the east and over 100 ft in the west.(Rickard, 1969).

In the major salt bed, mined at Retsof, New York, by the International Salt Company, the salt shows indistinct, coarse layering, without intercalating anhydrite or shale laminae

laminae found in other areas and units. These ill-defined, thick salt layers (1-3 ft), alternating only with each other and separable only through more or less sudden changes in the clay and organic content, resemble in many ways the crude layering of well-sorted beach-sands. This similarity is further enhanced by the presence of shale balls, 1-30 cm across, occurring in distinct horizons and traceable through much of the mine. Apparently, these shale balls, first noted by Dellwig (1969) were formed through avalanche-like growth of clay pebbles in the agitated shallow water, very much like the balls formed in loose volcanic tuff during rain. Thus they are remnants of ephemeral shale laminae, generally preserved in most sections of the Salina Group, but here washed away by entering waves. (It is interesting to note, that the short time available to Professor Dellwig in the mine disturbed his sedimentological interpretations. Strongly dipping crack systems in the shale underlying the salt, caused by blasting, were interpreted by him (1969) as a disconformity, with horizontal salt layers transgressively overlying the strongly dipping, dislocated shale. Following this line of thought, the shale balls appeared to him pebbles and cobbles of the abraded shale. Similar blasting cracks in the salt, on the other hand, were described and photographed by him as cross-bedding. Detailed study in the mine and in adjacent wells re-established the essentially parallel lamination in the salt and shale beds; though cross-bedding is theoretically possible, it could not be verified.)

According to Rickard (1969), the salt beds of Unit B are traceable through the Chatham Sag from the Appalachian into the Michigan Basin.

Upper Vernon (Unit C)

The undifferentiated red shales of the upper Vernon near the orogenic zone grade to the west into gray and green shales with interbedded siltstones, reaching a maximum thicknesss of over 400 ft. Farther to the west, dolomites become predominant; they contain anhydrite interbeds in northern Ohio. The thickness of the unit is gradually decreasing toward the platform zone and the arch system; separation of the northern Ohio subbasin is no longer evident.

In the Watkins Glen, New York area, only the uppermost beds of the upper Vernon are exposed by the wells. This top portion consists of green shale, grading into grey shale in its top 1 ft, where it underlies the Syracuse salt. The green shale is very silty and somewhat laminated. Laminae of more and less cemented (mainly anhydritic) shale alternate; intercalating silt laminae often display ripple marks. Silt lenses are also common. Finely crystalline anhydrite lenses, mostly a few mm across, occur frequently throughout.

The shale is impregnated with halite, which also fills its fissures. Thin beds of (and interstratal fissures filled with) red halite occasionally occur. The halite also forms lenses and nodules, few cm across.

Near the top, anhydrite becomes increasingly common. At first, it forms only a separating thin network between small lenses of halite, then separate lenses of anhydrite become larger in the shale matrix. Finally, about 0.2 ft below the bottom of the lowermost halite layer of the Syracuse formation, the top of a thin red halite bed is lined with a thin lamina of anhydrite. Although here anhydrite is dominant, the structure is still lenticular: a thin shale network separating the lenses makes up a minor portion of the lamina. The Syracuse salt does not follow directly: a thin shale bed with a few anhydrite lenses is intercalated.

Syracuse Formation

Apart from the relatively small amount of evaporites deposited during middle Vernon times, most of the evaporites of the Appalachian Basin form part of the Syracuse Formation. The cyclic sequence of the formation will be described primarily as it is developed in the axial zone, where, at Watkins Glen, New York, continuous cores of the International Salt Co. penetrate the formation in its entire thickness. These observations will be supplemented by data from the platform zone of the basin, where, at Cleveland, upper Syracuse salt is being mined by the same company.

Lower Syracuse (Unit D)

The lower portion of the Syracuse formation consists of dolomites and shales with two or three salt beds. The influence of the receding Bloomsburg delta, supplying silt and clay, is still evident. Maximum thickness of the unit (175 ft) and maximum aggregate thickness of the salt beds (over 80 ft) are attained in south-central New York and north-central Pennsylvania (Rickard, 1969).

In the Watkins Glen area, the Lower Syracuse (Fig. 4) consists of three major halite layers, separated by beds of shale or shale alternating with halite. The sequence penetrated by the standard borehole (WG 44) there is presented below:

D3 salt		30.0 ft
D2/3	shale 4	3.2 ft
	salt 3	5.5 ft
	shale 3	2.6 ft
	salt 2	6.3 ft
	shale 2	6.9 ft
	salt 1	4.4 ft
	shale 1	2.8 ft
D2 salt		28.8 ft
D1/2	dolomite	2.5 ft
	shale	7.9 ft
	salt	3.2 ft
	shale	1.8 ft
D1 salt		17.8 ft

All these horizons, even the fourfold repetition of shale and halite beds between salt layers D2 and D3, are well correlated in boreholes through several miles.

The Lower Syracuse at Watkins Glen, New York

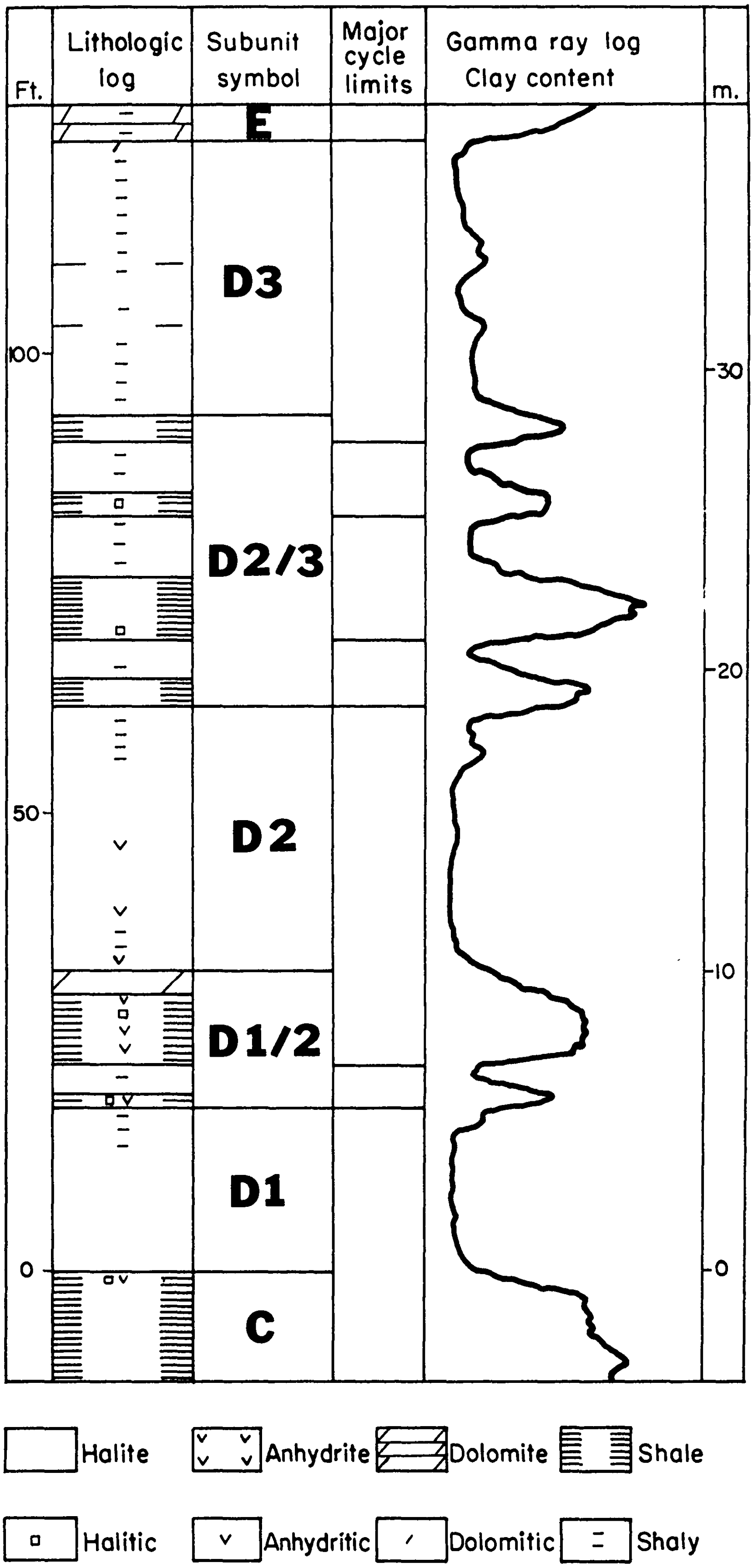


Fig. 4

D1 salt

The base of the first halite bed is uneven, like most shale-halite contacts of the Syracuse Formation. Salt crystals intrude into the underlying shale layer, resulting in a rugged contact. Presumably, this is not a primary sedimentary phenomenon; rather repeated minor recrystallization of the salt, accentuated at the shale-salt interfaces, and settling of the still plastic shale are the cause of the structure. The ruggedness of the contact seems to be independent of tectonic stresses, as it is characteristic also of undisturbed successions.

Above the contact, the lowermost 0.1 ft of the halite is still pink, as was the material of the salt lenses, fracture fillings and salt beds within the underlying green shale. Then, however, the halite turns gray, and remains so through the major part of the bed. The salt forms medium-sized crystals (4-10 mm) typical of the entire Syracuse Formation, while recrystallized larger clear crystals, occurring isolated or in aggregates, become increasingly numerous toward the central part of the bed (Fig. 5).

The top section of the bed, less than 5 ft thick, is more contaminated. First small relics of shale appear, concave-shaped, often sickle-like, then these join into a congruous network enclosing halite nodules. Two thin laminae of shale, 0.05-0.2 ft thick, intercalate in the salt. The lower, thicker one of these is gray, dolomitic, and contains frequent halite nodules. In spite of its relatively insignificant thickness, this lamina shows a typical

Figure 5

Medium grey, medium crystalline halite with frequent clear crystals.

Actual size

Lower Syracuse, Watkins Glen, New York

Figure 6

Shale lamina in salt, displaying small depositional cycle. Light grey anhydritic shale (S_A), containing pure white anhydrite lenses, forms loose network in the underlying halite and lines the base of the shale. Further up, the shale becomes almost anhydrite-free (S); salt nodules and crystals mark the anhydrite-free transition into the overlying salt (H).

Actual size

Lower Syracuse, Watkins Glen, New York

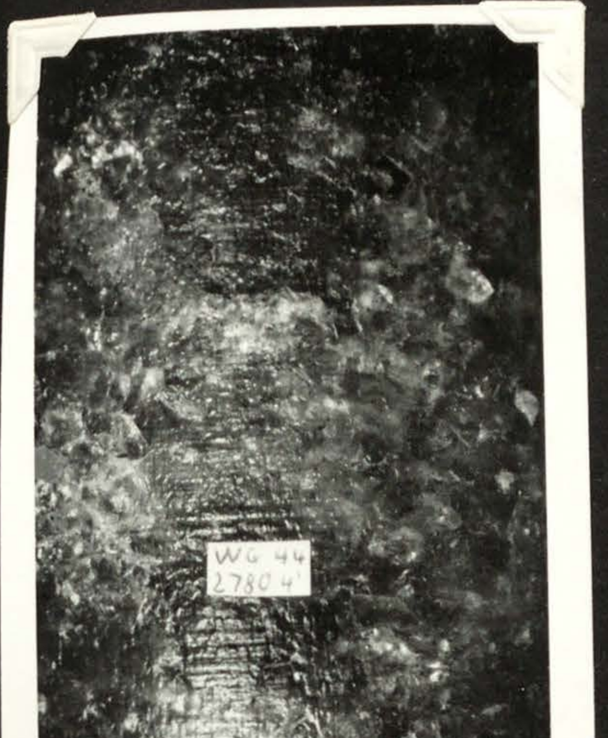


Fig. 5

Fig. 6

sequence (Figure 6): At its bottom, a lamina of shaly anhydrite lenses separates it from the underlying salt, further up only anhydritic shale concretions occur in the shale, near its top halite nodules appear, marking the transition toward the overlying salt. (Such halite-anhydrite-shale-halite cycles are frequent in the Lower Syracuse).

The upper shale lamina is thin, it consists of black shale with a few anhydrite lenses. Near the top of the salt pink colour returns; here irregular relics of greenish-grey shale are common.

D1/2 shale layer

Theshale above the lowest salt layer of the Syracuse resembles the Vernon shale in most respects. Like the thicker shale lamina in the D1 salt, it commences with a thin anhydrite zone (6 cm); the anhydrite is lenticular with the constituent lenses separated by a thin shale film. A relatively thin bed of shale follows, separated from the major part of the shale layer by a similarly thin salt interbed (3.2 ft). In spite of their small thickness, these beds are traceable over several miles.

This thin lower shale bed is greenish grey, well laminated; it contains frequent lenses of pure and shaly anhydrite and a few red crystals of halite. Upwards it becomes harder and more anhydritic, then grades into a lamina of uncemented pyritic clay. Here anhydrite is absent, while some red halite crystals still occur, possibly indicating a negative correlation between anhydrite and pyrite.

The salt interbed separating this lower shale bed from the major portion of the D1/2 shale layer is generally pink coloured. It contains a few relics of anhydritic shale near its bottom, followed by a thin shale lamina with relatively frequent anhydrite lenses.(Figure 7.). The bulk of the bed, however, consists of relatively shale-free greyish pink salt, with large amounts of pink to clear crystals dispersed singly and in agglomerates in the darker salt. Small relics of shale and anhydrite become numerous only near the top. Here, under ther microscope, wavy laminae of bituminous halite mark the layering in the salt, and systems of repeatedly bifurcating tubes, presumably branching algae, lie parallel to this lamination (Fig. 126).

The bulk of the shale layer overlies the halite interbed, again with anhydritic basal contact. Small pockets (Figure 8), possibly due to washout and leaching, deepen into the salt. These pockets are filled with fragments and lenses of anhydrite, embedded in an argillaceous anhydrite matrix, and grading upwards into a congruous lamina of shaly anhydrite containing very small pure anhydrite lenses. The entire anhydritic transition is about 6 cm thick.

Figure 7

Branching thin anhydritic shale lamina, containing frequent white anhydrite lenses, in salt

Actual size

Lower Syracuse, Watkins Glen, New York

Figure 8

Anhydrite (A) base of a shale bed overlying halite (H). The anhydrite fills irregular solution troughs in the salt with fragmented lenses randomly oriented; further up, the anhydrite lenses lie horizontally and decrease in number in the shale (S).

Actual size

Lower Syracuse, Watkins Glen, New York

Fig. 7

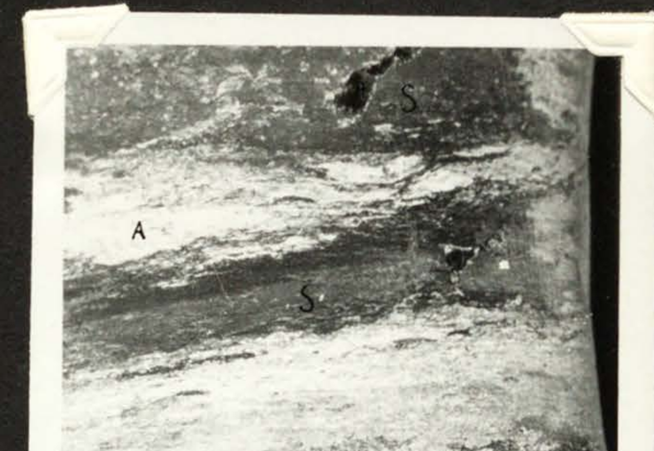


Fig. 8

Above its anhydritic basal contact, the major part of the D1/2 shale layer is greenish grey to greenish buff coloured; a few thin black shale laminae occur at its bottom only. Anhydrite lenses, on the other hand, are common throughout the shale. They often display a multiple-lenticular structure, in that small anhydrite lenses, with little shale intervening, associate to form larger lenses, and these, being more common in some zones than in others, associate to form lenticular laminae with "wiremesh" structure (Figures 9,10,11). Photo 10 shows the relationship of the anhydrite in the shale to the anhydrite contained in the halite interbeds: while in the underlying salt the anhydrite forms irregular, concave-shaped relics, in the overlying shale, as a rule, convex anhydrite lenses and concretions occur.

In its uppermost portion (3.6 ft) the shale becomes laminated and better cemented. At first the cement is still anhydritic, gradually, however, anhydrite gives place to dolomite and the rock grades into a typical biostrome. (Figure 12). The core of the hemispheroids building up the biostrome (see later) contains several salt-filled cavities; the algal crust displays alternation of irregularly undulating dolomite laminae with oscillating bitumen content, permeated and partially replaced by halite. The lamination is often interrupted by vertical structures. (Figure 13).

Upwards the cavernous biostromal laminae become gradually thicker (Figure 14). The dolomite becomes more halitic and it is

Figure 9

Anhydrite (A) zone with "wiremesh" structure in shale (S). The zone consists of closely packed anhydrite lenses with serrate margins separated by shale film.

Actual size

Lower Syracuse, Watkins Glen, New York

Figure 10

Anhydritic base of shale (S) overlying orange-coloured salt (H). The shale contains large upward convex anhydrite nodule with lenticular internal structure (A); the underlying salt contains small anhydrite relics.

Actual size

Lower Syracuse, Watkins Glen, New York

Figure 11

Coalescing anhydrite lenses displaying "wiremesh" structure in the basal part of a shale layer. The underlying salt contains small anhydrite relics.

Actual size

Lower Syracuse, Watkins Glen, New York

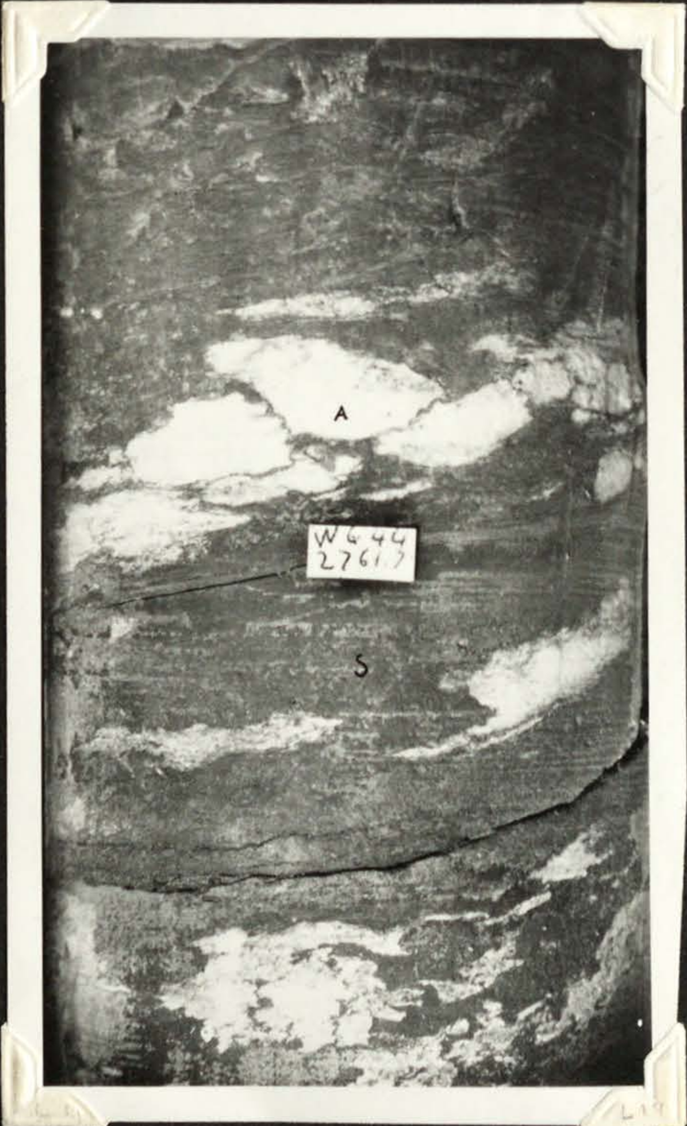


Fig. 9

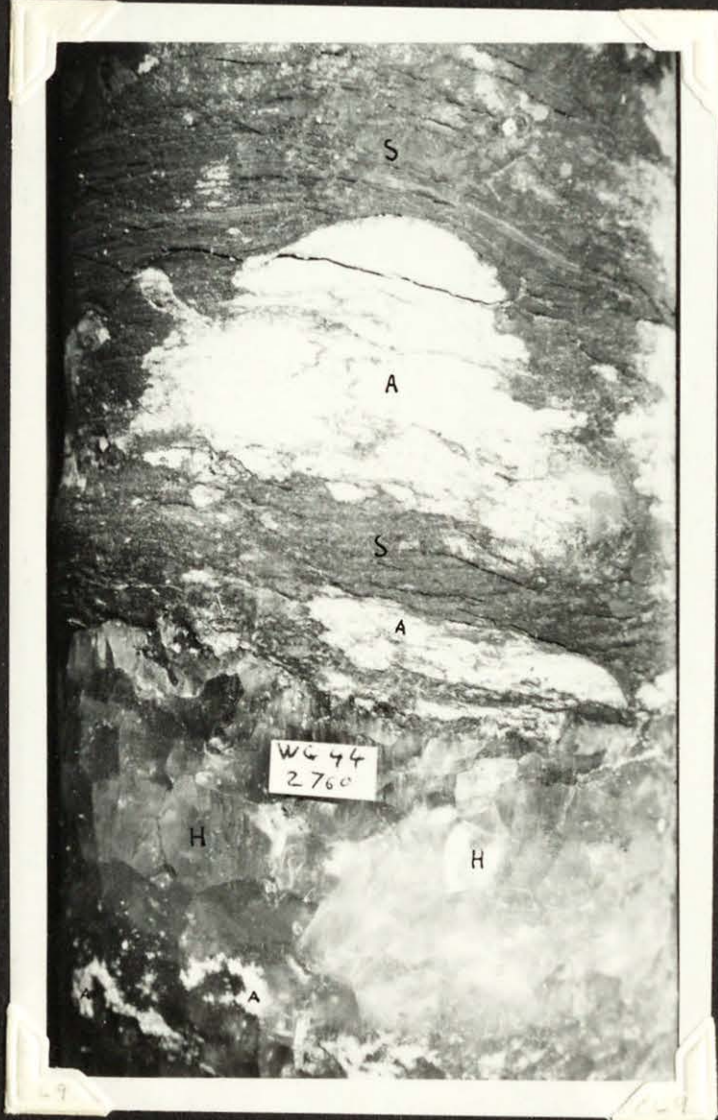


Fig. 10

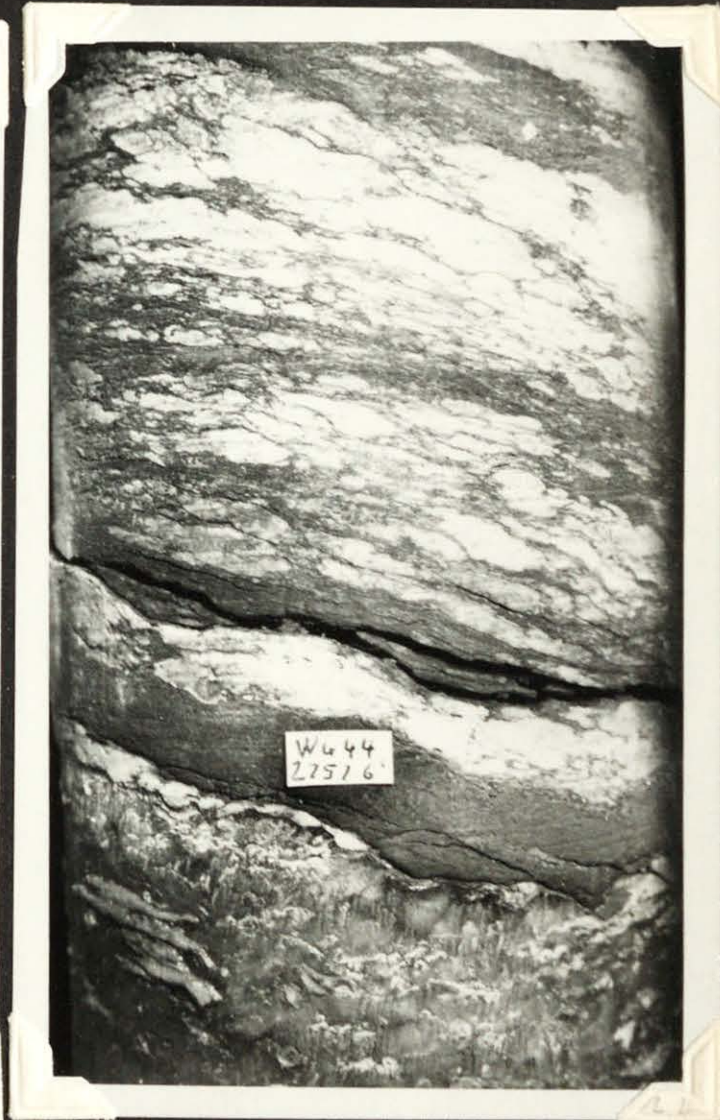


Fig. 11

Figure 12

Finely laminated algal biostrome. A hemispheroidal structure (bottom) grades upwards into fine, undulating laminae, in which dark, vertical bands (right centre) denote partial metasomatism of dolomite by halite. Cavities are filled with dark salt.

Actual size

Lower Syracuse, Watkins Glen, New York

Figure 13

Finely laminated dolomite biostrome. Cavities and vertical fissures are filled with dark salt and fringed with light, non-laminar dolomite.

Actual size

Lower Syracuse, Watkins Glen, New York

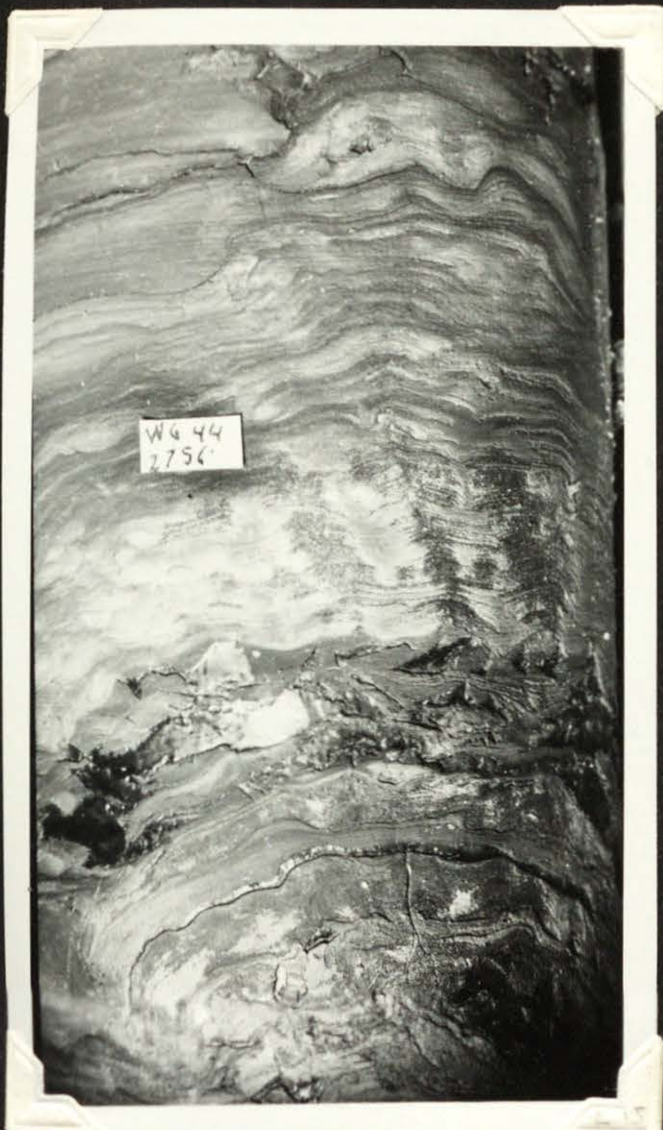


Fig. 12

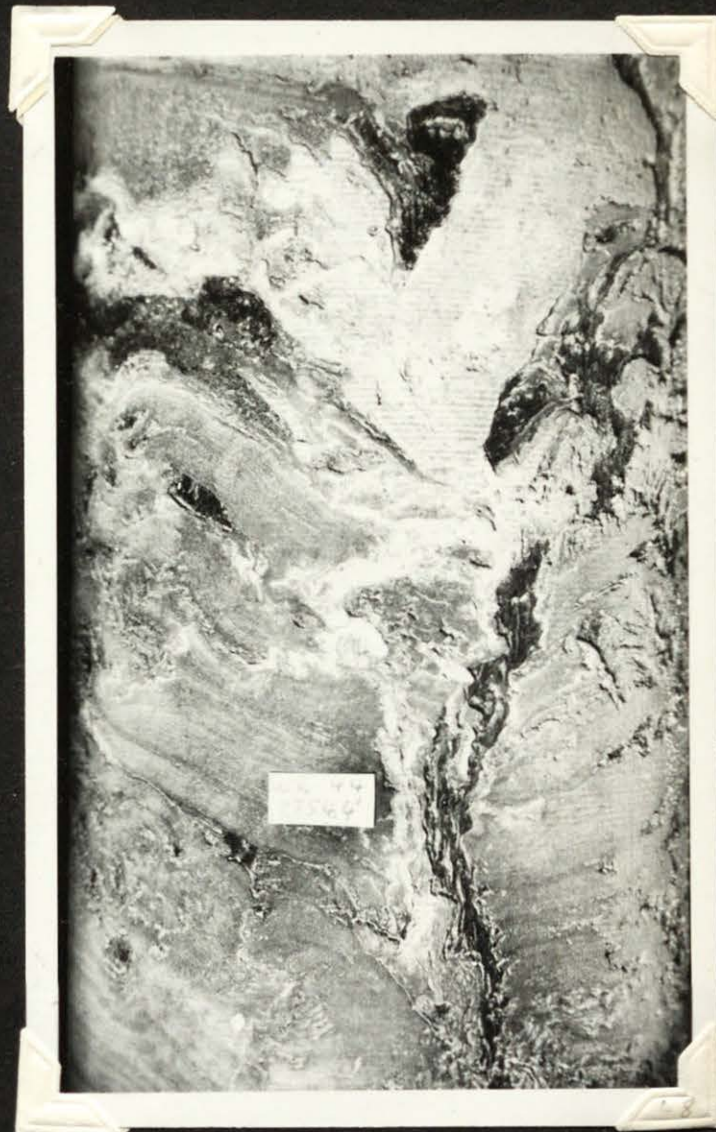


Fig. 13

ultimately overlain, along a comparatively flat contact, by the coarse, pale orange salt of the second halite layer (Figure 15).

Figure 14

Coarsely laminated algal stromatolite; the cavities are filled with dark halite. Along salt-filled fissure at right edge, the dolomite loses its laminar structure.

Actual size

Lower Syracuse, Watkins Glen, New York

Figure 15

Dark, massive dolomite, often pyritic, overlying the algal stromatolite of Fig. 14, and overlain by dark halite (H). Vertical cracks are filled with salt.

Actual size

Lower Syracuse, Watkins Glen, New York

Fig. 14

Fig. 15

D2 salt

The second salt layer resembles the first one in most respects. Pinkish at its bottom contact, it soon becomes light, rarely medium, grey. Dolomitic shale relics are rare, particularly in the lowermost 2 ft. Here, as well as in the lightest sections further up, a thin anhydrite film occurs between the halite crystals.

The upper 10 ft of the layer is again pink to pinkish grey, and contains abundant soft shale relics (a possible cause of the pink coloration).A few shale laminae, 1-2 cm thick, also occur (Figure 16). In the latter, the upper contacts have narrow depressions, filled with salt, while their lower contacts are saw-shaped and bear the imprint of the underlying salt crystals.

D2/3 shale and salt

The second and third salt layers of the Lower Syracuse are separated by a sequence of four shale beds, separated by halite interbeds. The basal contact of the lowermost shale (Figure 17) is almost completely flat. The shale beds are green and often display alternating, sometimes lenticular, laminae, which differ in their dolomite and/or silt content. Anhydrite lenses are generally absent, they occur only at the bottom of the fourth (and occasionally at the top of the third)shale bed.

Figure 16

Shale lamina (S) in halite (H). Irregular depressions at the top of the shale are filled with dark salt.

Actual size

Lower Syracuse, Watkins Glen, New York

Figure 17

Shale (S) overlying halite (H) along flat basal contact

Actual size

Lower Syracuse, Watkins Glen, New York

Fig. 16

Fig. 17

The halite interbeds are pink at their bottom and top contacts, their central part being grey. They contain abundant concave-shaped relics and a few irregular laminae (Figure 18) of soft geenish grey shale; the laminae often branch into a thick network separating crystals and nodules of salt (Figure 19).

The top contacts of the salt interbeds are frequently gradational, as the shale network becomes gradually predominant (Figure 20). In these cases, the lenticular lamination of the overlying shale is still disrupted by large salt nodules at its bottom.

Figure 18

Shale lamina (S) ramifying into discontinuous network in pink salt (H).

Actual size

Lower Syracuse, Watkins Glen, New York

Figure 19

Desintegrated shale network in salt

Actual size

Lower Syracuse, Watkins Glen, New York

Figure 20

Shale (S) overlying dark salt (H). In the top of the salt, shale forms discontinuous laminae and irregular relics; in the base of the shale, lamination is disrupted by salt intruding from below.

Actual size

Lower Syracuse, Watkins Glen, New York

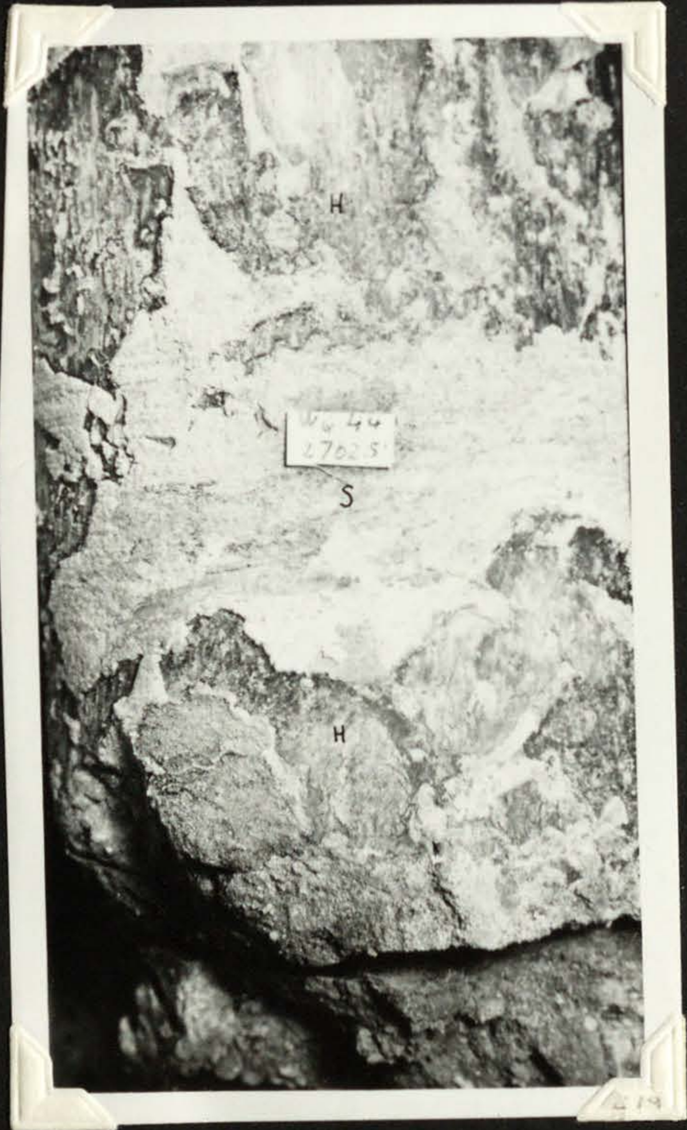


Fig. 18

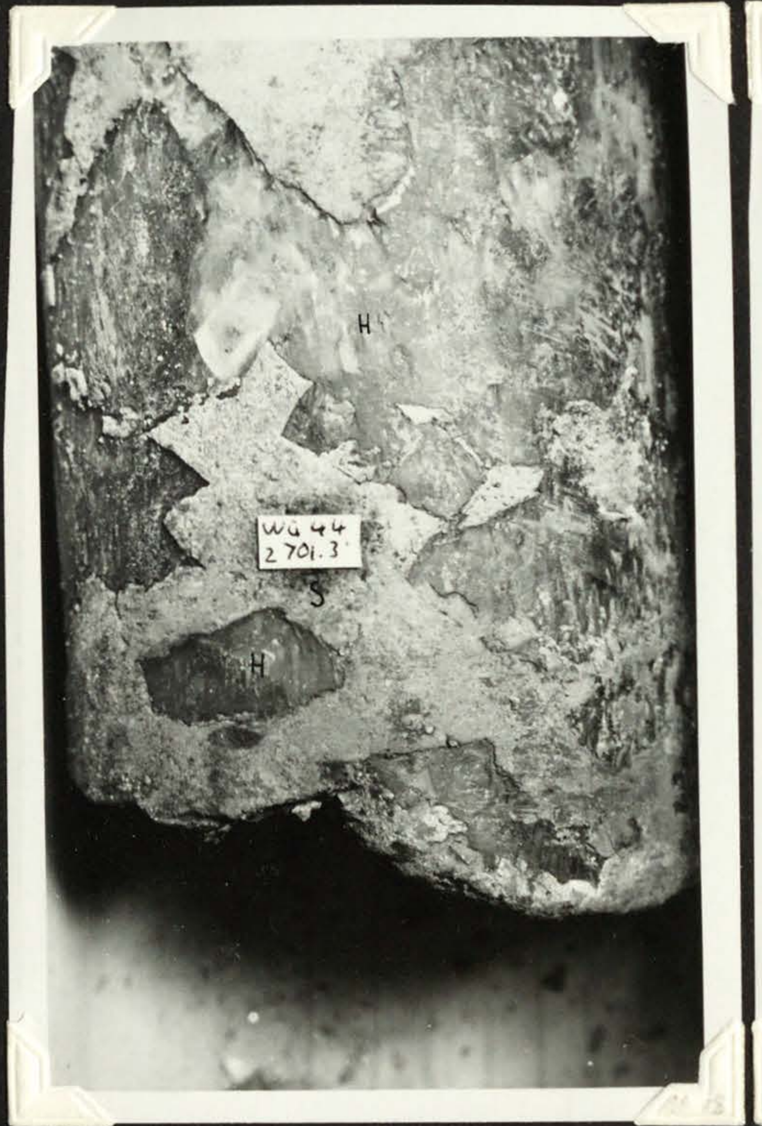


Fig. 19

Fig. 20

D3 salt

The third halite layer of the Lower Syracuse is about twice as thick as the lower two. It overlies the uppermost D2/3 shale bed with a very uneven contact: red salt lenses and crystals, frequent in the top of the shale, make the contact indistinct. The bottom of the salt is also red, then it gradually turns into pinkish grey. The upper half of the salt layer (about 14 ft) is medium to dark grey.

The salt is characterized by abundant shale inclusions. Apart from concave-shaped relics, whose form is the negative of the surrounding salt crystals, a few dark grey, fissile shale laminae (6-12 mm) also occur, particularly in the lower section. Larger clear crystals are fairly common, their number increases around the larger shale relics and shale laminae, presumably due to recrystallization along the shale/salt interfaces. In the lower, reddish half of the salt layer, these clear crystals also have a pinkish hue; in the upper, grey portion they are colourless.

In the top 2 ft of the salt layer, the shale relics are increasingly replaced by relics of argillaceous dolomite. The dolomite relics often form a loosely connected network. In sharp contrast to the lower salt beds which overlain by green shales showed the return of the pink colour at their top contacts, the upper contact of the D3 halite, overlain by reducing dolomitic-argillaceous rocks, is grey.

MIDDLE SYRACUSE (UNIT E)

The predominantly halitic Lower Syracuse Formation is overlain by a thicker sequence (about 150 ft) in which halite is subordinate. A lower, thicker dolomite suite (approximately 100 ft) and an overlying thinner shale suite (about 50 ft) can be distinguished (Fig. 21).

DOLOMITE SUITE (Subunits E1, E2, E3)

In contrast to the Lower and Upper Syracuse (Units D and F) in which the salt -shale-salt cycles are symmetrical, reflected in sinusoidal gamma-ray and neutron-gamma curves, cycles of the Dolomite Suite are strongly asymmetrical both in the Appalachian and in the Michigan Basins. Sudden appearance of shales and /or very argillaceous dolomites over virtually clay-free rocks is followed by a gradual decrease of clay content until it reaches its previous low level. The most frequent cycle is

Halitic dolomite
Argillacous, frequently anhydritic dolomite
Argillaceous dolomite and shale
- -
Halitic dolomite

Beds of pure halite, separated from the underlying dolomite by a thin band of anhydrite, occasionally occur, giving a composite cycle of

Halite
Anhydrite
Dolomite
Shale

The Middle Syracuse at Watkins Glen, New York

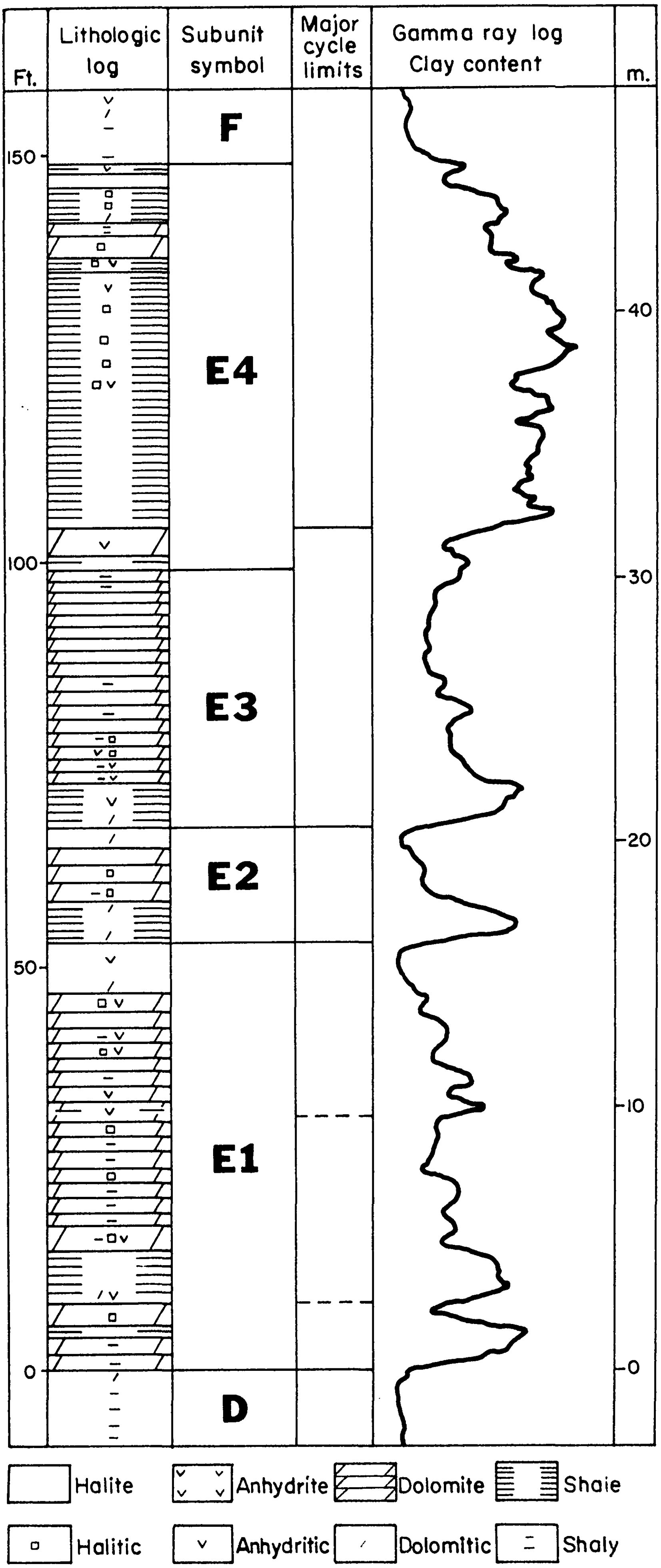


Fig. 21

While in this latter presentation the sequence is typical of evaporite deposits, in the Middle Syracuse Dolomite Suite shale, anhydrite and halite occur mostly in the from of interbeds, laminae, or accessory components, and the sequence consists primarily of dolomite. Salt layers comprise less than 10% of this unit.

On the basis of its oscillating clay content, the Dolomite Suite can be subdivided into three major subunits, of which Subunits E1 and E2 are strongly, E3 is slightly asymmetrical (Fig. 21). The essential sequence is presented below:

E3	Halitic dolomite	20.2 ft
	Argillaceous dolomite	6.4 ft
	Shale and arg. dolomite	5.2 ft
E2	Halite	2.5 ft
	Halitic dolomite	6.9 ft
	Shale and arg. dolomite	4.7 ft
E1	Halite	6.4 ft
	Halitic dolomite	23.5 ft
	Argillaceous dolomite	8.9 ft
	Shale and dolomite	14.8 ft

Subunit E1

Basal shale and dolomite

The most argillaceous basal members of the E1 subunit consist of two shale layers, separated by a dolomite interbed. The lower shale layer overlies the top of the Lower Syracuse salt with a thin transitional sequence. At the top of the salt,halite nodules, 2-5 cm large, are contained in a network of mostly vertical dolomitic shale fragments. In the overlying dolomite lamina, only small lenses of halite and anhydrite crystals are present, then gradual incrase of clay content leads into the lower, greenish grey shale layer. The structure of this increasingly shaly interval is lenticular-laminated: the amount of purer flat dolomite lenses decreases while that of the wavy shale laminae amoung them increases. Some of the purer dolomite lenses and laminae persist, however, even in the purest shale, though due to their rarity the rock here becomes more massive. The thin dolomite laminae and lenses become more frequent upwards, towards the dolomite interbed; the dark thin argillaceous network amoung them turns dark grey to black and is often pyritic.

The intercalating halitic dolomite will be discussed later in greater detail. It is of interest, however, to point at a thin (6 cm) anhydritic boundary between the dolomite, and the overlying upper shale layer. This boundary zone consists of alternating thicker anhydrite and thinner argillaceous dolomite laminae, and displays an interesting structure on wet surface."Ghosts" of monoclinic crystals (Figure 22) appear in random position,vertical

Figure 22

Anhydritic boundary zone between halitic dolomite (D) and overlying shale (S). The zone consists of alternating dark argillaceous dolomite (D_S) and white anhydrite (A) laminae. On the wet surface, "ghosts" of anhydrite pseudomorphs after large gypsum crystals (G) can be discerned; they cut across the lamination and seem to make up part of the anhydrite laminae.

Actual size

Middle Syracuse, Watkins Glen, New York

Figure 23

Photomicrograph of argillaceous dolomite, consisting of thicker laminae of argillaceous dolomite with abundant light anhydrite lenses and thin laminae of dark argillaceous-carbonaceous matter.

Magnification 12 diameters

Middle Syracuse, Watkins Glen, New York

Figure 24

Intraformational dolomite breccia in darker argillaceous matrix (S) underlain by argillaceous dolomite containing brecciated as well as intact dolomite laminae.

Actual size

Middle Syracuse, Watkins Glen, New York

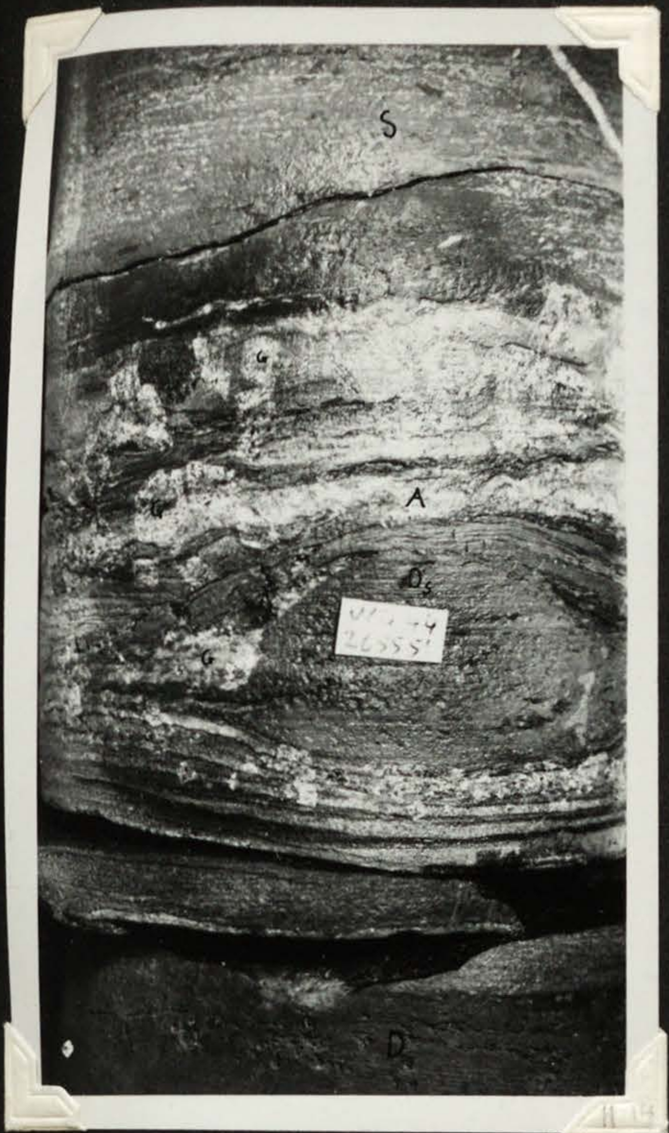


Fig. 22

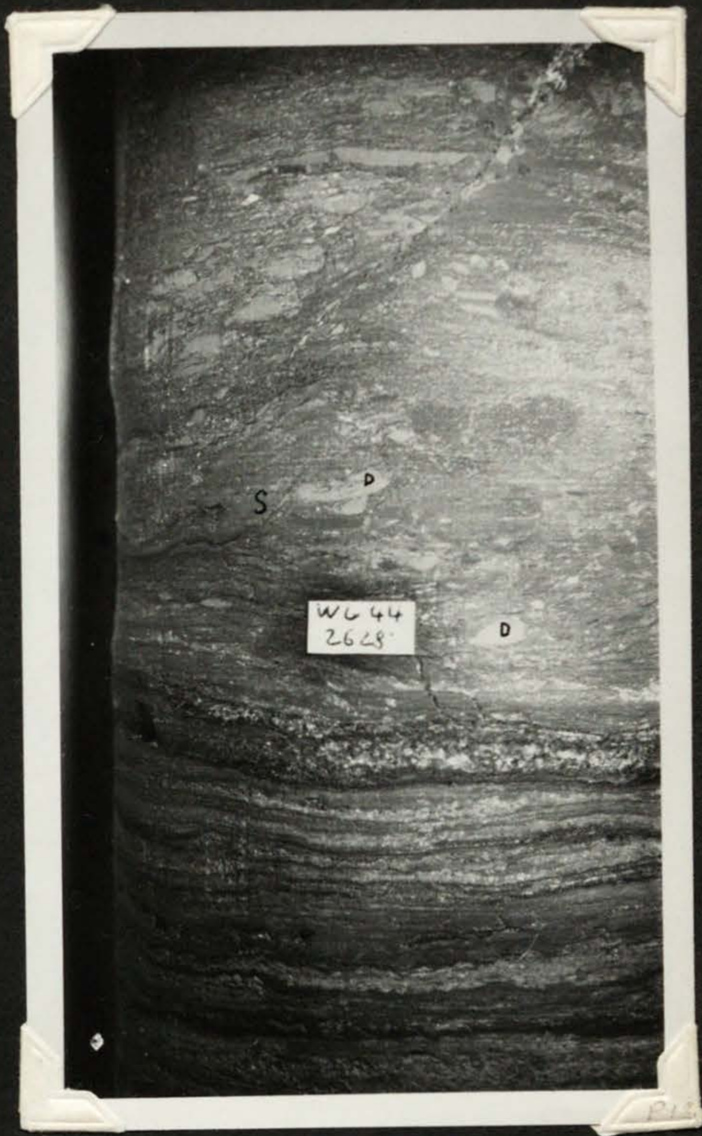


Fig. 24

Fig. 23

to horizontal, often deeply intruding into the argillaceous dolomite laminae.Presumably, they are pseudomorphs after randomly oriented gypsum crystals formed during the early diagenesis of the sediments and replaced by anhydrite at a later stage.

Argillaceous dolomites

Alternating layers of halitic and slightly argillaceous dolomite form the major part of the E1 cycle. The clay content decreases in steps from each major argillaceous to each major halitic dolomite layer (2-6 ft each). The generally low clay content of these halitic dolomite interbeds also decreases gradually towards the main, thick halitic dolomite layer. Two such cycles of oscillatingly decreasing clay content can be distinguished in Subunit E1.

The argillaceous dolomite layers normally display microscopic lamination. Thinner laminae of strongly argillaceous-carbonaceous dolomite and shale, often containing finely crystalline pyrite, alternate with thicker laminae of clay-free dolomite (Figure 23). Small lenses of anhydrite are common, their centre (1-2 mm) is occasionally occupied by halite. Less frequently, the dark argillaceous-shaly laminae are slightly wavy, and the dolomite forms long thin lenses between them. If the clay content is low, only thin wavy lenses of argillaceous substance occur.

Some sections display intraformational breccia:slightly rounded dolomite fragments are embedded in more argillaceous matrix (Figure 24).

In zones grading upwards from argillaceous into halitic dolomite, (Figure 25) the wavy -lenticular lamination is often disturbed by incipient halite crystallization. Where, on the other hand, halitic dolomite is overlain by strongly argillaceous dolomite or shale, intraformational breccias displaying slight rounding occur.

Halitic dolomites

The lowest layer of halitic dolomite occurs, as mentioned above, intercalated in the basal dolomitic shales. Here the dolomite is greyish brown; halite is contained in the form of larger crystals, lenses and spherules. The halite lenses, 3x10 mm near the base, gradually disappear, and the dolomite becomes saturated with small spherules of salt (0.3-3mm), containing finely dispersed brown do= lomite. Some dolomite laminae are "consumed" by such dolomitic halite spherules and salt crystals, so that only a discontinuous cellular dolomite network remains (Figure 26). In contrast to the dolomite network in the purer halite layers, here the halite lenses and laminae never exceed a few mm in thickness, are never transparent but always strongly dolomitic, and most of the halite bodies are only a few mm in diameter. The larger dolomitic halite bodies never form continuous laminae or beds, but wedge out in the dolomite matrix within a few dm. A semi-laminar structure is, however, sometimes perceptible, as the pellets and lenses of dolomitic halite are concentrated in some dolomite laminae and absent in others.

The halitic dolomite layers above the basal shales are simpler

Figure 25

Dolomite with wavy argillaceous laminae(D_S) grading upwards into dolomite (D) containing dark halite lenses.

Actual size

Middle Syracuse, Watkins Glen, New York

Figure 26

Dolomite (D) partially permeated and replaced by halite (H_D). Pure, light dolomite dominates at bottom and top, but it is preserved only in the form of discontinuous relics in the middle. At top, dolomitic halite forms zones of small spherules in the dolomite; at bottom, short steep cracks, some of them displaying feather-like structure, may be of desiccational origin.

Actual size

Middle Syracuse, Watkins Glen, New York

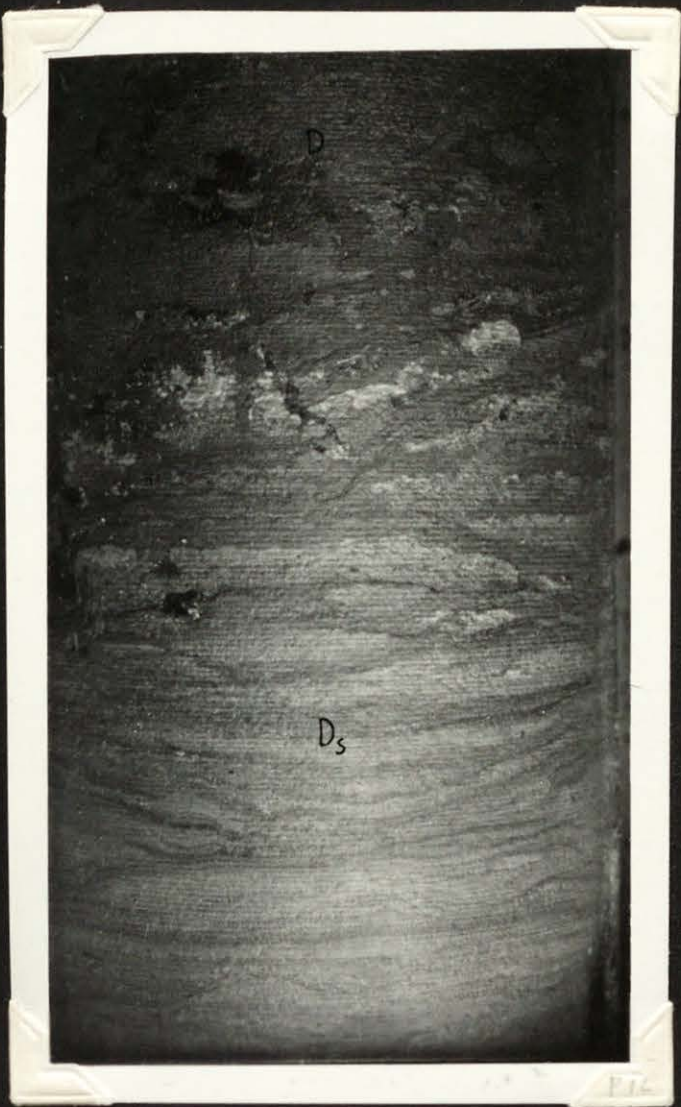


Fig. 25

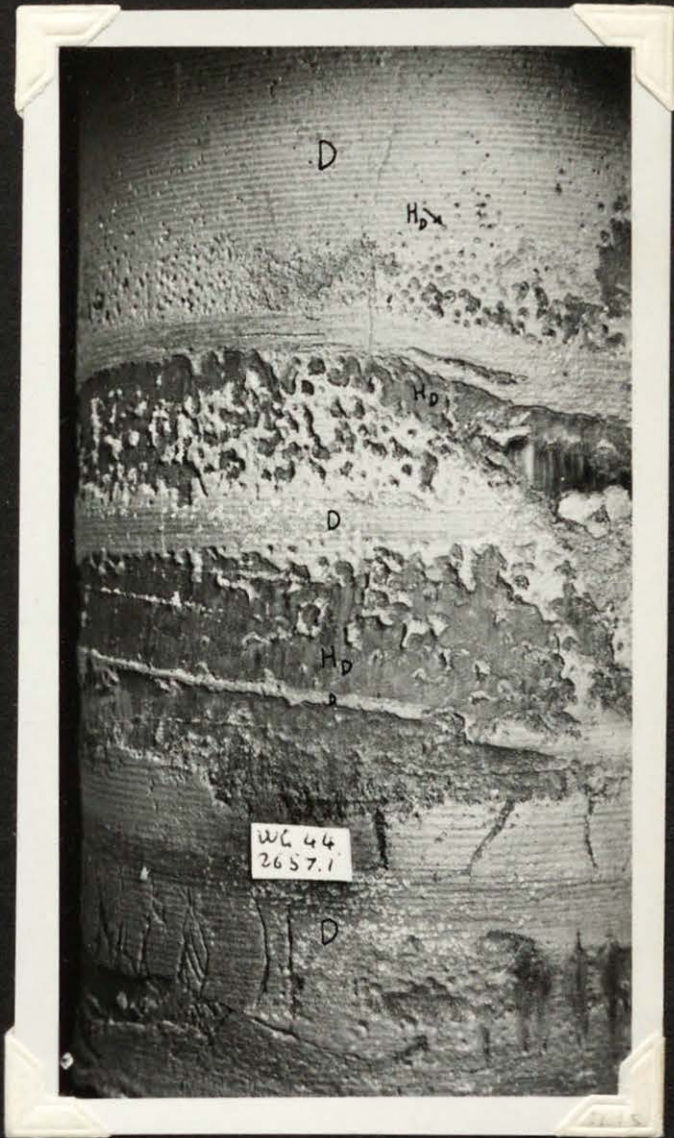


Fig. 26

in structure. Salt spherules occur less often, the salt more frequently forms lenses and crystals (up to 1 cm) ; a few discontinuous thin salt laminae also occur. Some of the latter have a pyritic film along their contact with the enclosing dolomite. Upwards in the halitic layer, these larger lenses and thin laminae (0.3-2 mm) are alternately rich and poor in halite spherules. Locally, thin laminae of halitic dolomite (1-2 mm) alternate with thicker laminae of halite-free dolomite. Beds of halite-free dolomite contain pyrite. Often, the top contacts of the halite-free dolomite layers are sharp; above such contacts,apparently erosional, dolomitic halite spherules are very common. From a few fissures, intricate networks of rootlike microfissures, filled with halite, ramify upwards and connect some of the halite spherules (Figure 27).

In the top 1.2 ft below the E1 salt, the structure of the dolomite somewhat changes. Overlying wavy, apparently erosional surfaces, thin laminae of coarse dolomite silt (0.1 mm) intercalate. They soon give place to thin laminae of anhydrite (0.4-4 mm) which contain small halite crystals.

Halite

The uppermost member of the E1 subunit is halite. At its base, laminae of shale and anhydrite are common; the shale laminae contain thin lenses of shaly anhydrite and are bordered by an anhydritic margin. This anhydritic material continues into the intercrystalline spaces of the adjacent salt laminae, giving them a cellular structure (Figure 28). Above this base, light and medium grey salt beds alternate, both lighter than the salt of the

Figure 27

Faulted surface of halite-free dolomite (D) overlain by halitic dolomite (D_H). The halitic dolomite forms intraformational breccia in darker, more halitic matrix at the bottom, and contains small patches and spherules of dark dolomitic halite (H_D) near top. Some of the latter are connected with dark halitic channel system, ramifying upwards from the fissure in the underlying pure dolomite.

Actual size

Middle Syracuse, Watkins Glen, New York

Figure 28

Fragments of dark shale lamina (S) reflecting flowage of salt (H). The light anhydritic rim of the shale fragments extends to form intercrystalline walls between adjacent salt crystals.

Actual size

Middle Syracuse, Watkins Glen, New York

Fig. 27

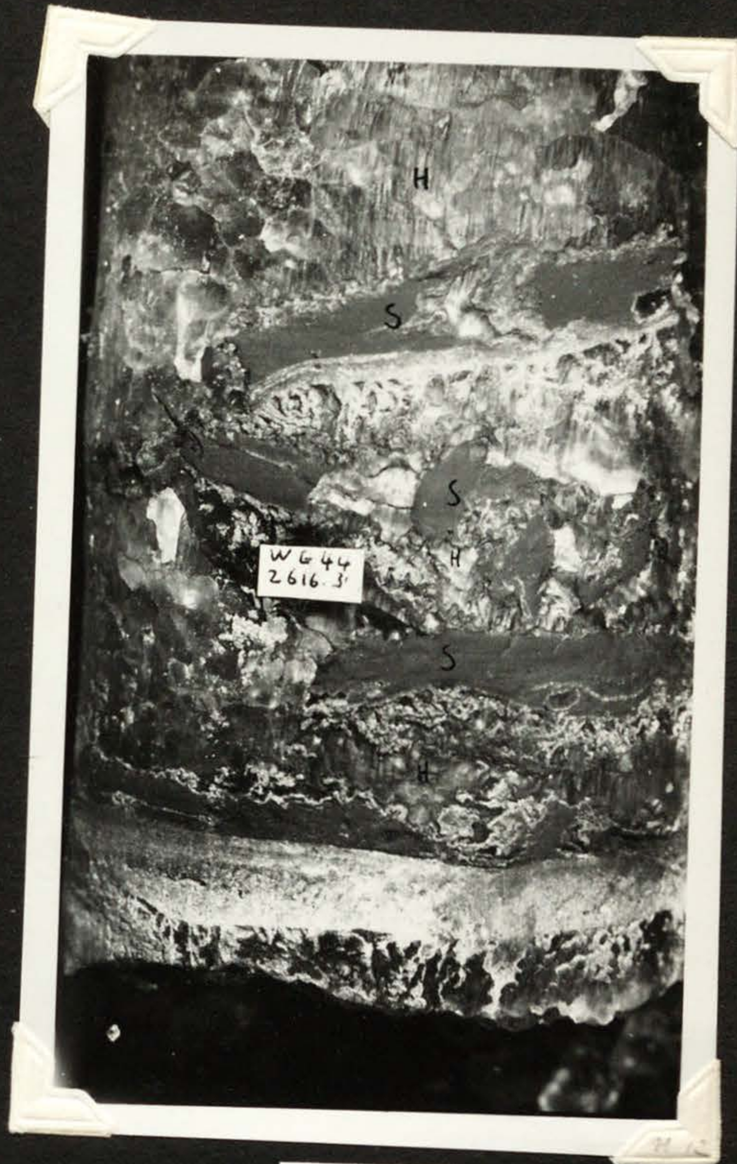


Fig. 28

more halitic Upper and Lower Syracuse. Particularly in the lighter sections, the salt crystals are coated by thin intercrystalline anhydrite film, while relics of dolomite are virtually absent.

Subunit E2

Subunit E2 repeats the E1 cycle on a somewhat smaller scale. The basal shale grades upwards into less and less aregillaceous dolomite and finally into halite.

In the basal shales, some beds contain intraformational breccias. Slightly rounded fragments of pure dolomite, 1-3 cm large (Figure 29) are randomly embedded in more argillaceous matrix. Otherwise slight lamination is produced by the oscillation of the dolomite content; a few lenses of anhydrite occur throughout.

The overlying dolomite is very halitic. Spherules of dolomitic halite (less than 1 mm) are very common (Figure 30); lenses of salt occur frequently. The lenses are often several cm long and several mm thick; they follow the intricately curved structures of bioconstructed (mainly algal) encrustations. Near the top, salt bands of constant thickness (1cm or less) occur. Above them the dolomite is penetrated by a network of mudcracks, filled with salt (Figure 32).

The higher salt content of the E2 dolomites is reflected in the presence of a halite interbed (0.35 ft). It develops gradually from the underlying biostromal dolomite, which contains halite - filled cavities increasing in size upwards.(Figure 31). Eventually, only flat relics of dolomite remain, several cm long, lying in

Figure 29

Intraformational dolomite breccia. Slightly rounded dolomite fragments are embedded in dark argillaceous matrix.

Actual size

Middle Syracuse, Watkins Glen, New York

Figure 30

Halitic dolomite. The halite, generally dolomitic, forms zones of small spherules and larger, halite-permeated dolomite patches, merging near top into a semi-continuous lamina. Steep discontinuous cracks are filled with halite.

Actual size

Middle Syracuse, Watkins Glen, New York

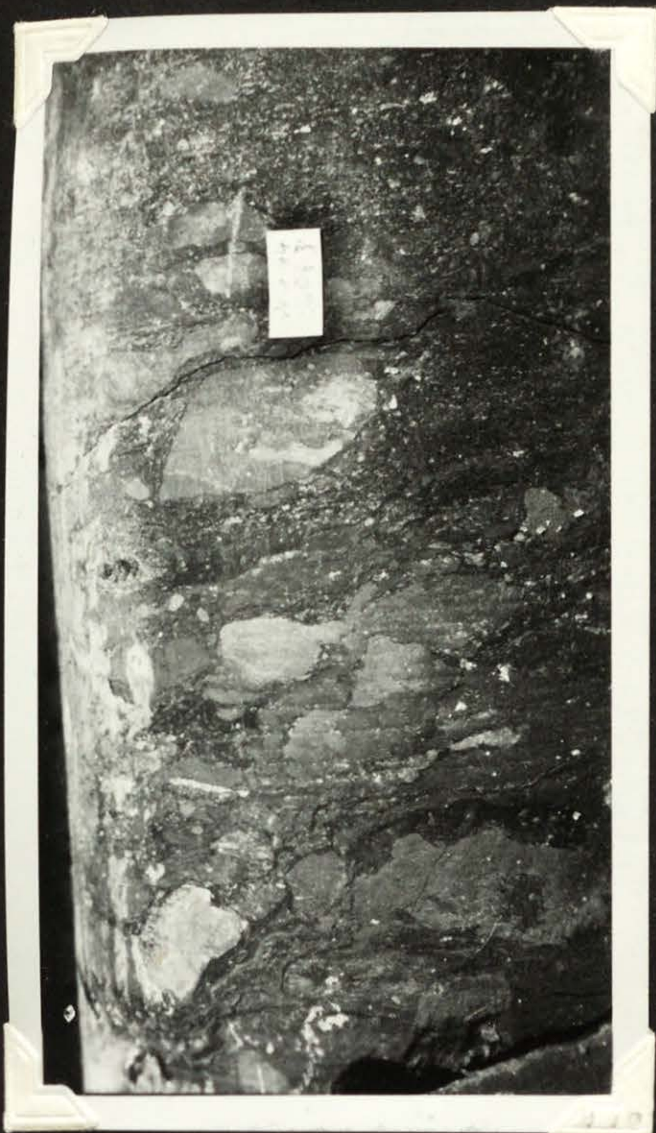


Fig. 29

Fig. 30

Figure 31

Dolomite reef-rock; round cavities are filled with dark halite

Actual size

Middle Syracuse, Watkins Glen, New York

Figure 32

Slightly laminated dolomite displaying desiccational (?) structure. The crack system is filled with halite. Near top, two bright halite laminae intercalate.

Middle Syracuse, Watkins Glen, New York

Fig. 31

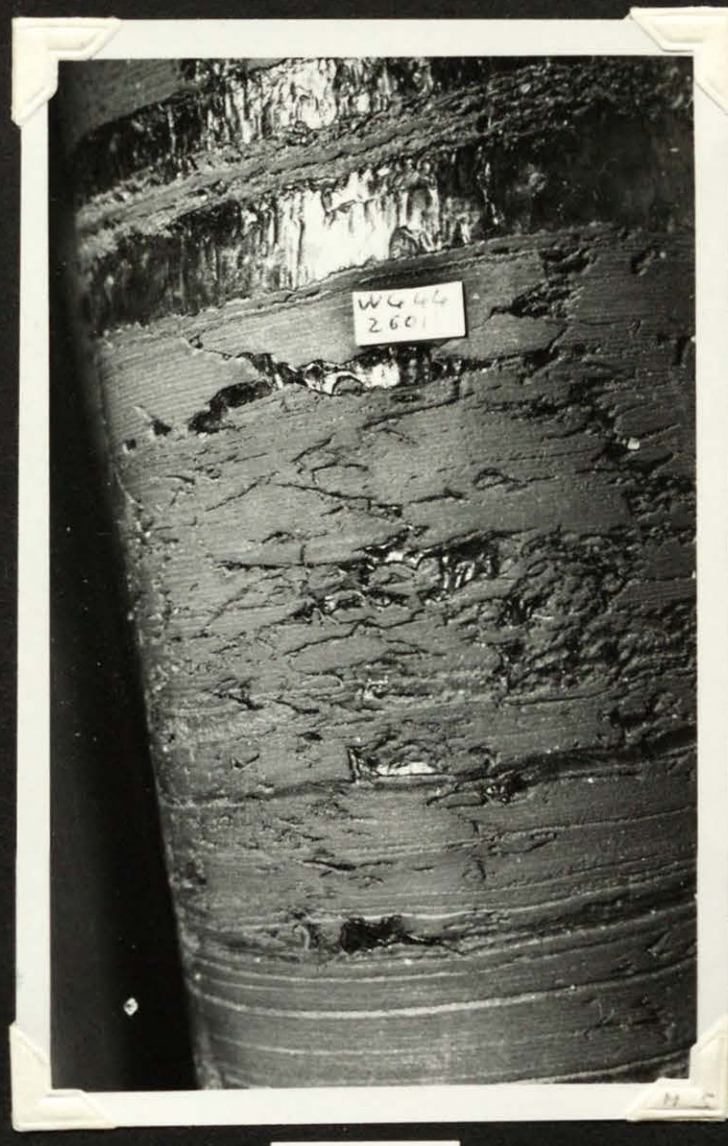


Fig. 32

random positions in the dominant halite.

Like Subunit E1, E2 ends with a halite layer. The major, lower part of the halite is light gray. Broken, dislocated dolomite laminae near the base of the salt show minor flowage. Further up, a few thin interbeds of laminated dark argillaceous dolomite occur; they contain lenses and crystals of medium gray salt. In the clear to light gray salt beds, intercrystalline anhydrite film is common. Near the top, the salt turns medium gray, and the small dolomite fragments are replaced by more shaly relics.

Subunit E3

The third subunit of the Dolomite Suite differs from the lower ones in the absence of separate halite layers at its top. Here the basal shales are overlain by coarse dolomite, displaying varied structural features and grading upwards into the Middle Syracuse Shale Suite.

Basal Shales

The basal shales of the subunit overlie the E2 salt with a slightly wavy contact (Fig. 23). The contact is lined with anhydrite, 2-3 cm thick; it consists of two laminae of almost pure anhydrite, separated by an argillaceous anhydrite lamina of similar thickness. The internal lenticular structure of the anhydrite laminae is vaguely perceptible; a few small, flat halite lenses also occur.

The lower part of the basal shale is dolomitic and well laminated. Near its base, the laminae are almost flat. Small anhydrite lenses are common, they occasionally join into strongly plicated laminae. As the clay content increases upwards, the anhydrite laminae become increasingly distorted or irregular; black shaly partings, similarly plicated, parallel them. Near the top, the clay content of the shale increases and the lamination disappears.

Argillaceous dolomite

In the argillaceous dolomite overlying the basal shale,

<u>Figure 33</u>

Anhydritic boundary zone (A) between halite (H) and overlying shale (S). The anhydrite fills shallow depressions in the salt and contains a somewhat darker argillaceous anhydrite lamina.

Actual size

Middle Syracuse, Watkins Glen, New York

<u>Figure 34</u>

Weathered surface of massive dolomite, permeated by superficial cracks and covered by dolomite fragments, overlain by laminated argillaceous dolomite containing bands of rounded dolomite fragments.

Actual size

Middle Syracuse, Watkins Glen, New York

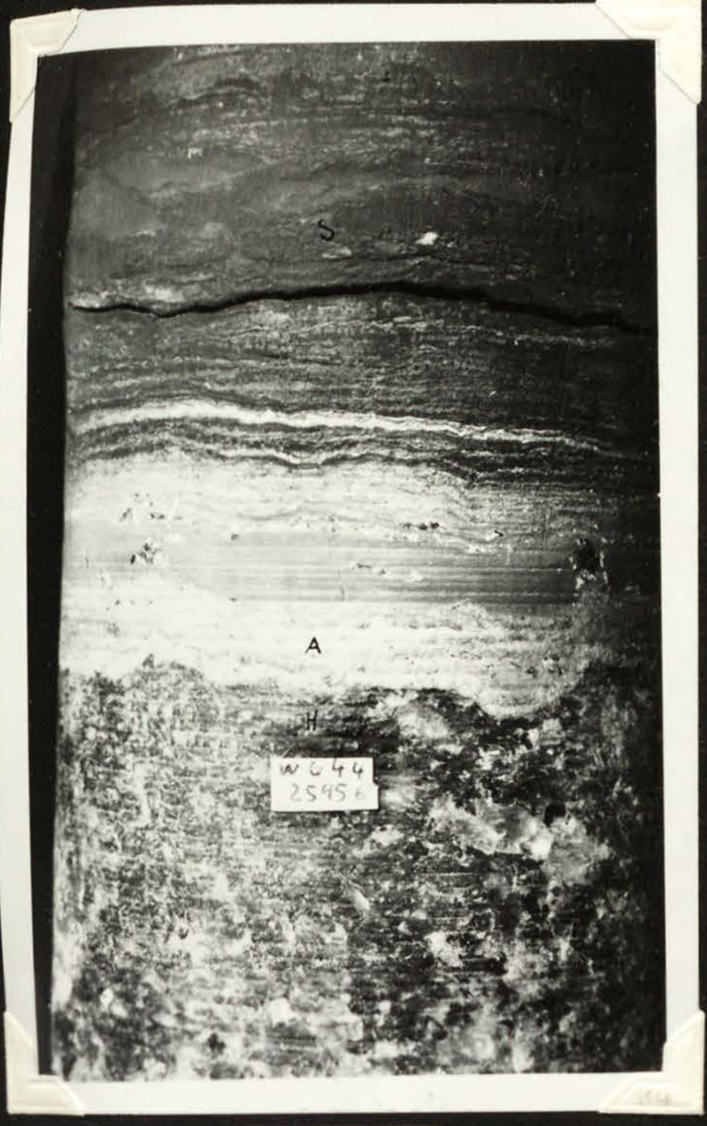


Fig. 33

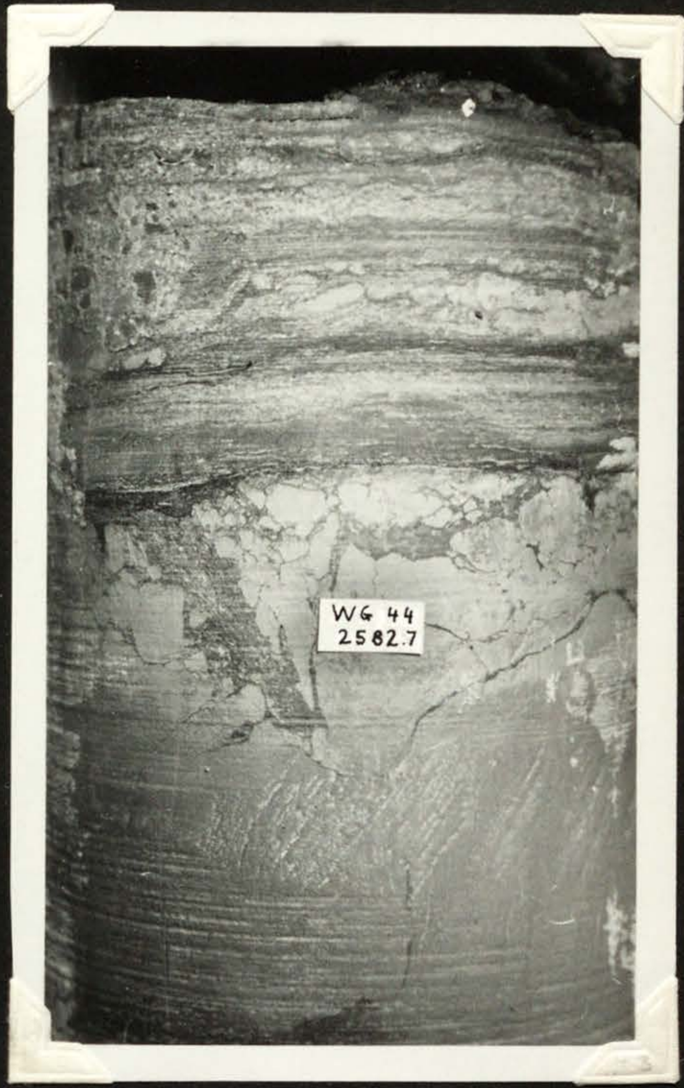


Fig. 34

flat shale laminae are still common, often producing fine lamination. Except for the very bottom, salt is a common constituent. First, it forms very thin (-1 mm) long lenses and thin laminae. Then somewhat thicker laminae (3-5 mm) become frequent, they consist of small salt crystals. The top portion is again virtually salt-free. Here, between an argillaceous and a pure bed of dolomite, an erosional surface occurs (Figure 34). From this surface, numerous fractures proceed a few cm deep into an underlying massive dolomite bed, desintegrated into small fragments in the vicinity of the erosional contact. Above it, alternating laminae of more and less argillaceous dolomite follow, incorporating small fragments of argillaceous and pure dolomite.

Clay-poor dolomite

The relatively clay-free upper portion of the dolomite displays complex biostromal structures. Most of the dolomite shows the fine lamination of algal-stromatoporoidal incrustations; the alternation of dark grey and lighter brown laminae reflects the oscillation of organic matter. A few beds of coarse dolomite silt intercalate. Near base, the laminae are slightly undulating, further up they are flat; some of the partings are partially coated with fine pyrite.

The coarser dolomite, absent in most of the Salina Group, consists of well-sorted rounded dolomite rhombohedra, 0.03-0.05 mm across. These grains are about an order of magnitude larger than the general grainsize of the dolomite rocks; with them, a few quartz grains of similar to slightly smaller size also occur.

Figure 35

Bioconstructed cavernous dolomite lens in massive dolomite; pores and cavities are filled with dark salt

Actual size

Middle Syracuse, Watkins Glen, New York

Figure 36

Bioconstructed cavernous dolomite lamina in massive dolomite; pores and cavities are filled with dark salt

Actual size

Middle Syracuse, Watkins Glen, New York

Fig. 35

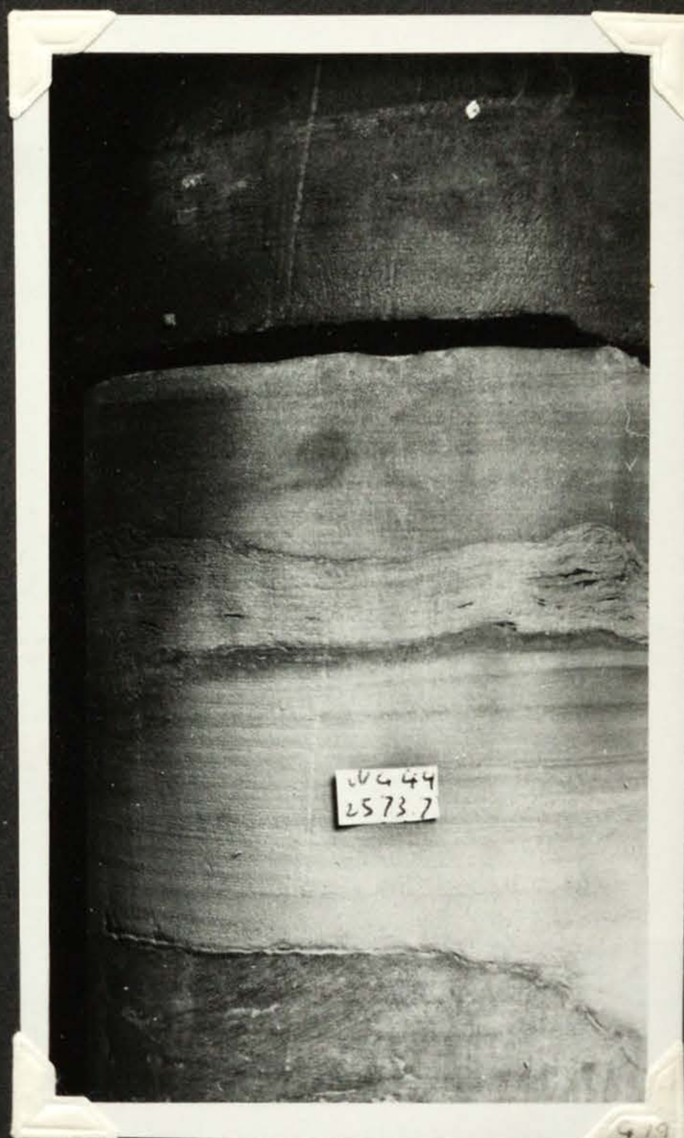


Fig. 36

Single distinct laminae (Figure 36) and lenses (Figure 35) of cavernous, bioconstructed dolomite often occur in the massive dolomite layers; the cavities, often typical of stromatoporoids, are filled with salt which also often cements the dolomite grains. Locally, isolated dolomite rhombohedra occur in the locally dominant halite matrix. Together with the salt lenses, a few anhydrite needles (0.3 mm) appear in random position, incorporating and cutting across dolomite grains; irregular discontinuous short fissures filled with halite and dark organic matter are common.

The coarse dolomite beds normally overlie the finer-grained beds at very sharp, erosional contacts. Frequently, however, they also form lenses and laminae in the fine-grained matrix. In evaporite-free interlaminated sets, burrow-like structures can be observed (Figure 37); they intrude up to 10 cm deep into the underlying bed and are filled with coarse dolomite.

These interlaminated beds often show intricate structures: somewhat disturbed sets of laminated dolomite are more or less incorporated in the dark, fine-grained massive dolomite (Figures 38, 39). Above and below such structures, the lamination is undisturbed. Apparently, the finely laminated bioconstructed sets suffered minor sliding in the fine-grained, argillaceous carbonate mud.

Figure 37

Light colomite grading upwards into finer-grained carbonaceous-argillaceous dolomite.

A steep burrow is filled with coarse dolomite silt.

Actual size

Middle Syracuse, Watkins Glen, New York

Figure 38-39

Finely laminated algal stromatolites wedging out in massive dolomite and displaying some slumping

Actual size

Middle Syracuse, Watkins Glen, New York

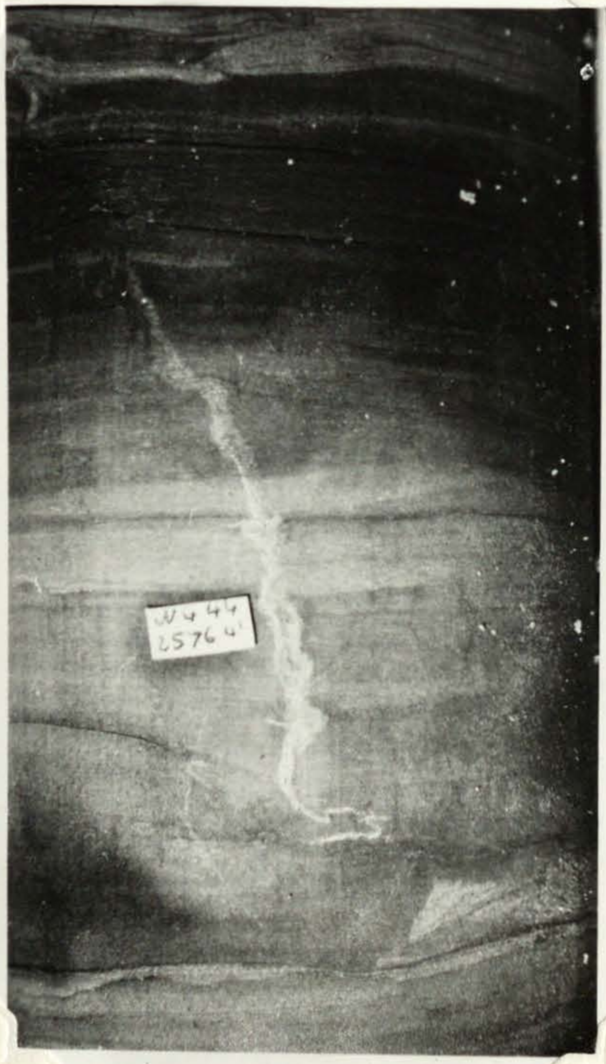


Fig. 37

Fig. 38

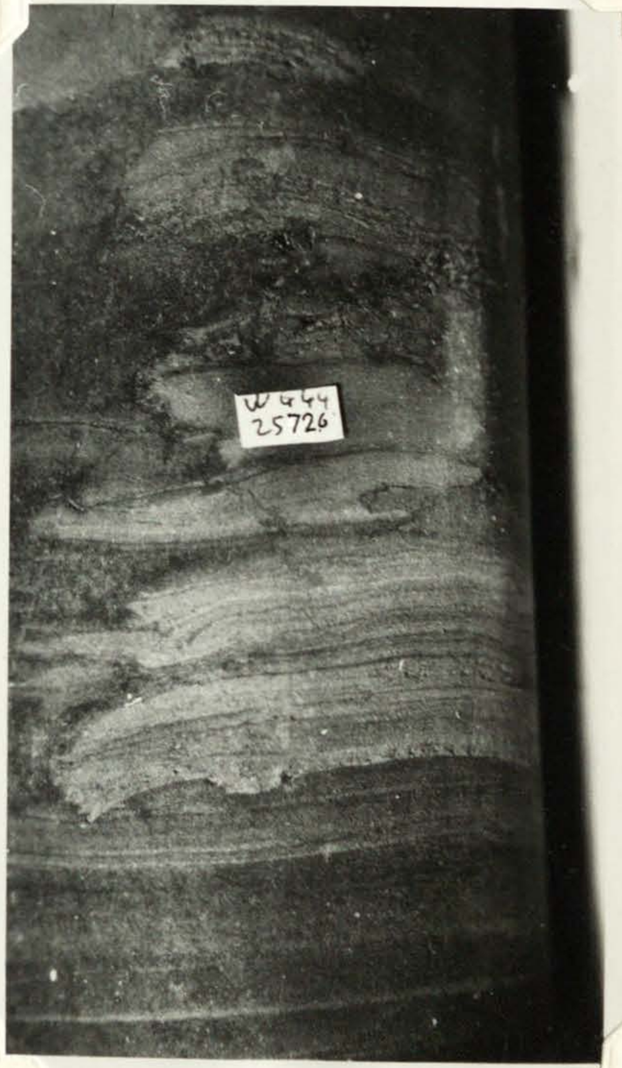


Fig. 39

SHALE SUITE (Subunit E4)

As opposed to the major, lower part of the Midddle Syracuse sequence, which consists chiefly of dolomite, its upper third consists mainly of shale. This thick shale layer contacts the underlying E3 dolomite as well as the overlying Upper Syracuse salt with chracteristic transitional sequences.

Basal transitional sequence

As in several instances before in the lower, more dolomitic part of Unit E, the boundary of the shale with the underlying (halitic) dolomitie is strongly anhydritic. A finely interlaminated set of dolomite, anhydrite, and shale occurs here (Figures 40, 41). The shale laminae often display minor undualations; the dolomite laminae frequently contain small lenses of halite. Of greater interst is a regular hexagonal system of mudcracks on the bedding planes; in vertical section they are particularly evident in the anhydrite laminae (Figure 40). Micro-unconformities are common, evened out by the overlying lamina.

Close to the bottom of the thick shale sequence, a dolomite interbed intercalates: it is finely laminated with alternating thinner argillaceous-bituminous and thicker, pure dolomite laminae (Figure 42). Lenses and crystals of anhydrite are common, the shape of the lenses occasionally resembles gypsum crystals. A few halite lenses also occur.

Apart from this relatively thin dolomite interbed, in the

<u>Figure 40, 41</u>

Basal anhydritic zone of shale. Dark shale and white anhydrite laminae alternate. Some anhydrite laminae display desiccation cracks and fill depressions in the underlying shale lamina; others contain small shale flakes.

Actual size

Middle Syracuse, Watkins Glen, New York

<u>Figure 42</u>

Anhydritic dolomite with white anhydrite lenses. The latter form cumulus-type aggregates and are accompanied by dark argillaceous laminae.

Actual size

Middle Syracuse, Watkins Glen, New York

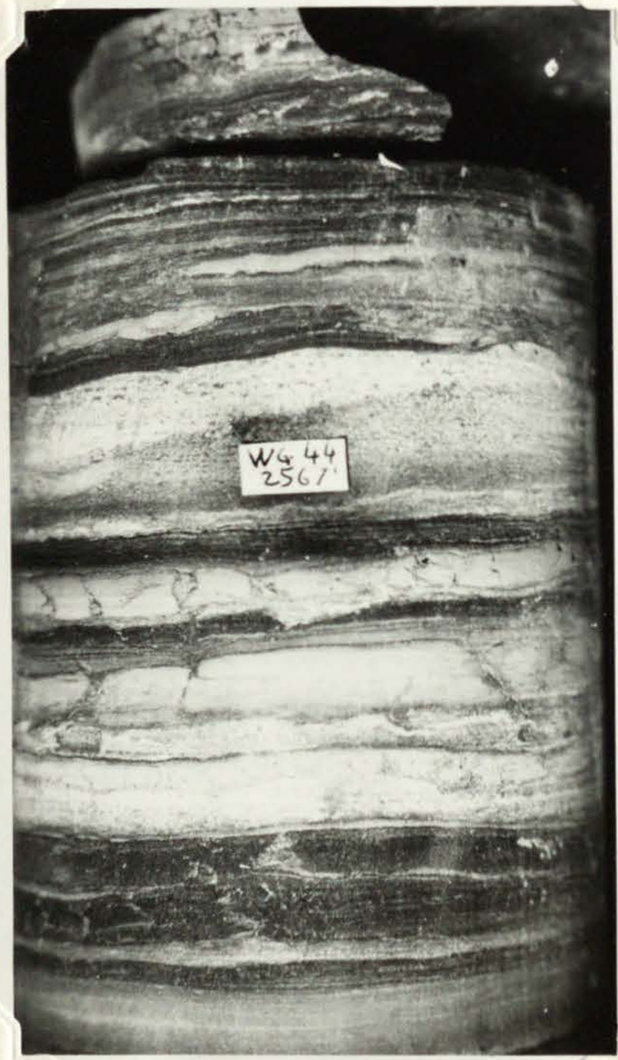


Fig. 40

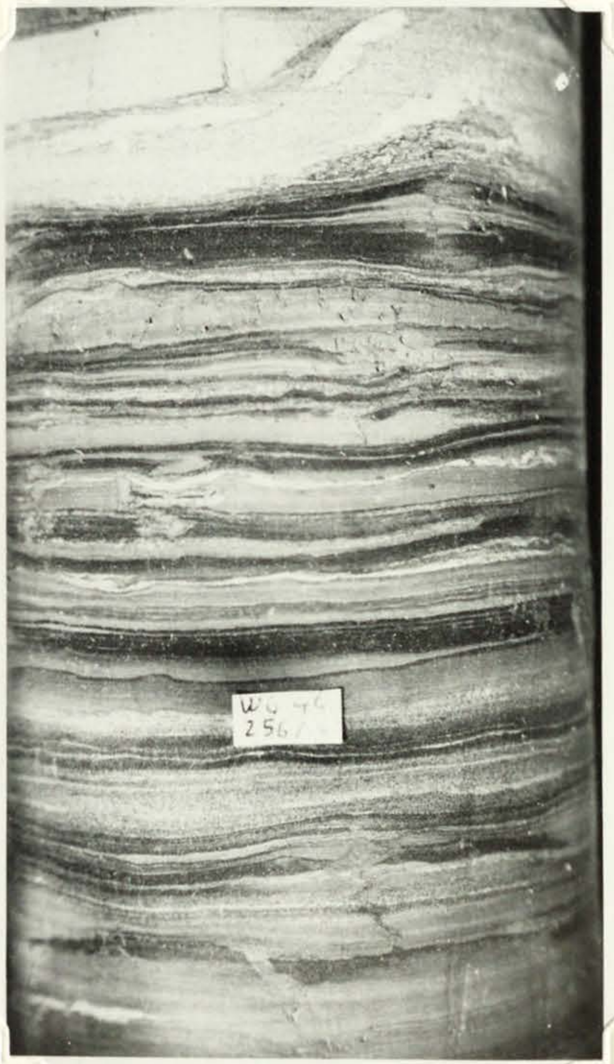


Fig. 41

Fig. 42

major, lower portion of the Shale Suite, shale is exclusive. In the upper 16 ft, a few beds of (or interstratal fissures filled with) pale orange to light yellowish gray halite intercalate. The salt/shale interfaces are sharp; the basal contacts are often lined with anhydrite. A few shale fragments occur in these salt beds; occasionally traces of fibrous structures, perpendicular to the shale/salt contacts, are perceptible.

Throughout the Shale Suite, the shale is medium to dark gray, somewhat dolomitic and intensively silty. Muscovitic quartz silt and fine sand (0.05-0.08 mm, rarely +0.3 mm) form flat lenses and flat or wavy laminae. Some laminae display cross-lamination under the microscope, with straight dolomitic shale and silt laminae alternating in the micro-foresets. Sometimes erosional channels are visible even in the silt-free laminae; in these, the base of the micro-channel fills contains more dark, bituminous material than their top section. Some of the silt-free laminae contain large amounts of silt-sized pyrite crystals.

Of particularly great interest are possible trace fossils. In a silt lamina (Fig. 43) a system of tubes, resembling a candelabrum, branches outwards and upwards from a central point, breaks through a shale lamina, and joins undulating 'burrows' above it.

A few thin beds (0.7 ft) consist of almost uncemented, soft greenish gray clay; generally, however, there is some cementing material, chiefly salt and dolomite, present.

Figure 43

Photomicrograph of a worm tube system penetrating a dark argillaceous-carbonaceous lamina in silty shale

Magnification: 12x

Middle Syracuse, Watkins Glen, New York

Figure 44

Star-shaped pressure cracks surrounding growing salt crystals in dolomite. Radii are filled with white anhydrite and recrystallized dolomite.

Actual size

Middle Syracuse, Watkins Glen, New York

Figure 45

Salt crystals growing in dolomite are surrounded and connected by cracks and channels filled with white anhydrite + recrystallized dolomite

Actual size

Middle Syracuse, Watkins Glen, New York

Small lenses of pale orange to light grey salt are common; from their central core, salt rootlets often ramify into the surrounding shale. Around single lenses and crystals, often short (1-3 cm), star-shaped, wedging out radial fissures are formed, normally at 120^{o} angles. While halite intrudes into these radii, they are mainly filled with recrystallized dolomite and anhydrite, surrounding the salt lenses. (Fig. 44). The recrystallized dolomite crystals contain relics of shale, crystals and relics of anhydrite, lenses of halite, and a few recrystallized crystals of quartz. Larger recrystallized anhydrite crystals occur less frequently, they contain halite lenses with anhydrite relics in their centre.

Less often, irregular, curved fissures connect the halite lenses; they often terminate abruptly and are also filled with recrystallized dolomite and anhydrite (Fig. 45). Sometimes, the anhydrite coating ~~around~~ the salt lenses unites to form a continuous lamina of anhydrite.

Thin fissures filled with pale orange salt are also lined with anhydrite and recrystallized dolomite, with minor amounts of quartz, all growing on the shale along the walls of the fissure, towards the salt-filled centre. Sometimes the fissures are discontinuous: small crystals and lenses of salt arranged along a steeply dipping plane are separated by intervals of fissure-free shale (Figure 46). The trace of such fissures is often very irregular, stylolite-like.

Apart from lining halite, anhydrite is rare in the lower part of the Shale Suite; it becomes progressively more common upwards. Near the top of the shale, strongly plicated anhydrite

Figure 46

Dolomite containing darker anhydrite concretions (A), halite crystals surrounded by radial pressure cracks, and discontinuous fissure filled with halite possibly pseudomorphous after gypsum

Actual size

Middle Syracuse, Watkins Glen, New York

Figure 47

Dolomite with large dark halite nodule at bottom and contorted light laminae at top.

Actual size

Middle Syracuse, Watkins Glen, New York

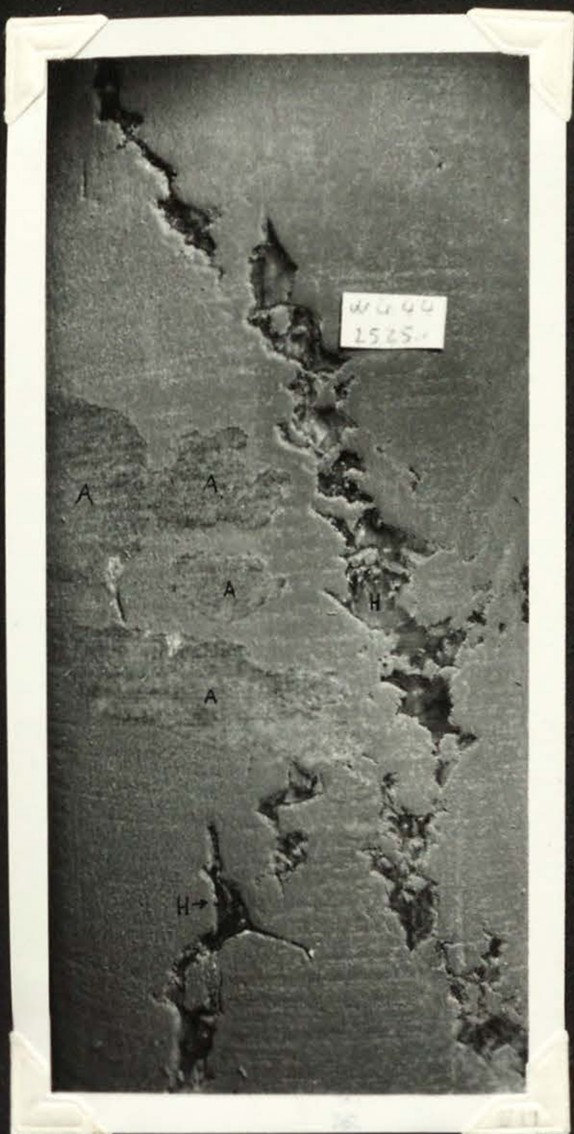


Fig. 46

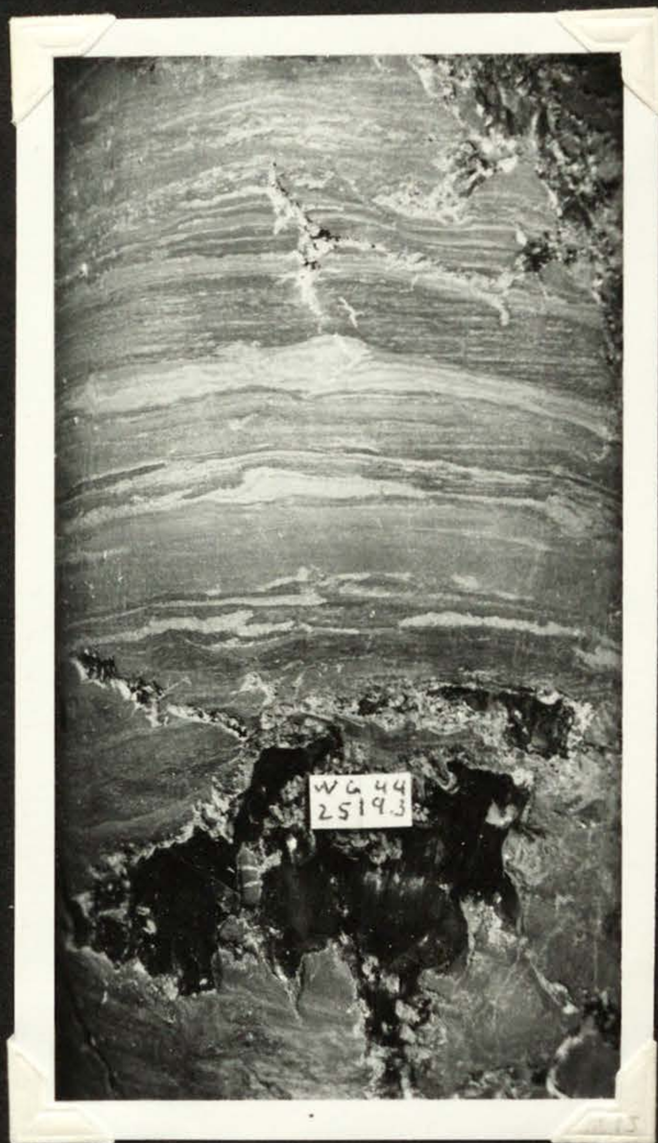


Fig. 47

laminae occur together with large, irregular halite nodules (Fig. 47). A few very thin shale laminae, enclosed in the anhydrite laminae, are intricately folded. These microscopic shale laminae are almost opaque in thin section, possibly from very fine pyrite; in reflected light, they are light to medium grey. Locally, they are lined by large, secondary dolomite crystals, growing into the anhydrite lens. Some of the anhydrite lenses are more than 5 cm large and often join to form discontinuous anhydrite laminae (Figure 48).

Close to the top of the shale, a thicker, somewhat coarsely crystalline (1 cm) interbed of salt is already grey, without the pink hue of the lower interbeds, and resembles the overlying Upper Syracuse salts. The bottom and the top of this bed are light grey, the central portion is darker, though clear lenses are common throughout.

Figure 48

White anhydrite lens between large dark salt nodules disrupting lamination in dolomite

Actual size

Middle Syracuse, Watkins Glen, New York

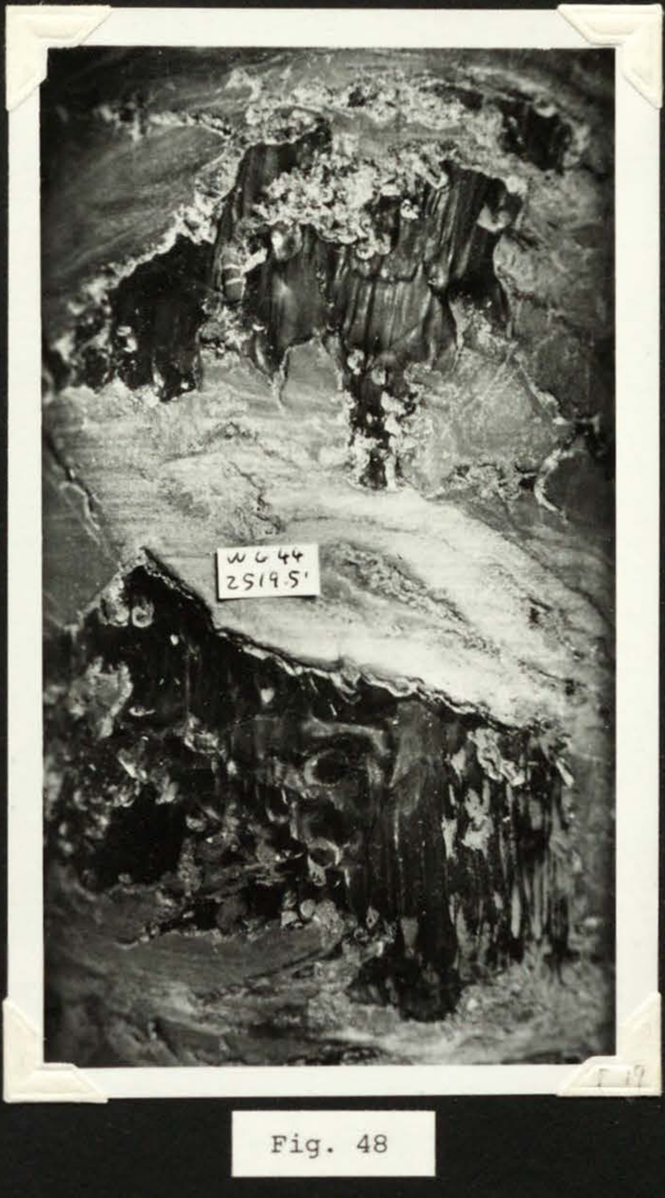


Fig. 48

Upper Syracuse (Unit F)

The upper Syracuse contains the majority of the salt that occurs in the Salina Group of the Appalachian Basin. The salt forms thick, dark, bituminous beds, contaminated with and separated by dolomite or shale. The shale fragments range from a few hundred microns to several cm, and display an intercrystalline relic structure: their shape is controlled by the surrounding halite crystals. The shale relics are often sparsely dispersed, discontinuous. Where they are more frequent, they join each other: first thin shale "threads" connect the larger relics, then the threads become thicker and a continuous network of up to 1 cm thick shale "threads" permeates the salt. Finally, the shale (or dolomite) becomes dominant, containing only irregular nodules of halite. Such oscillations, ranging from almost pure shale to halite containing only a few shale relics characterize the Upper Syracuse salts.

In the thick Upper Syracuse sequence, the following subunits are distinguished (Figures 49,50):

F3+4 salt	165 ft
F3/2 shale and dolomite	45 ft
F2 salt	50 ft
F2/1 shale and dolomite	53 ft
F1-2 salt	31 ft
F1-2/1 shale	11 ft
F1-1 salt	64 ft

F1-1 salt

The lowermost salt layer of the Upper Syracuse (Unit F) develops gradually from the Middle Syracuse Shale Suite. As mentioned above, in the top of the latter a thicker gray salt bed intercalates, whose characteristics already resemble those of the F salts. Above this salt layer, anhydrite lenses are frequent in the uppermost shale beds.

At its bottom, the Upper Syracuse salt, though grey, still shows some pinkish hue in its recrystallized clearer crystals. Shale relics are common and even a few intensely brecciated dark grey dolomitic shale laminae occur. Further up they give place to laminae and relics of dolomite (Figure 51), gradually ramifying without brecciation in the salt. The dolomite network is of particular interest (Figure 52): it consists of elements often having a sharp point downwards and flat surface on top, as if the dolomite filled the spaces between upward pointing halite crystals. Locally, even a bed of pure dolomite appears, having sharp bottom contact and grading upwards gradually into salt with increasingly frequent large salt nodules (5 cm). Concave-shaped dolomite relics are common, even within the nodules. Under the microscope, the dolomite is coarsely crystalline, it consists of slightly rounded large rhombohedra like those constituting the top of the Middle

Upper Syracuse, bottom section at Watkins Glen, New York

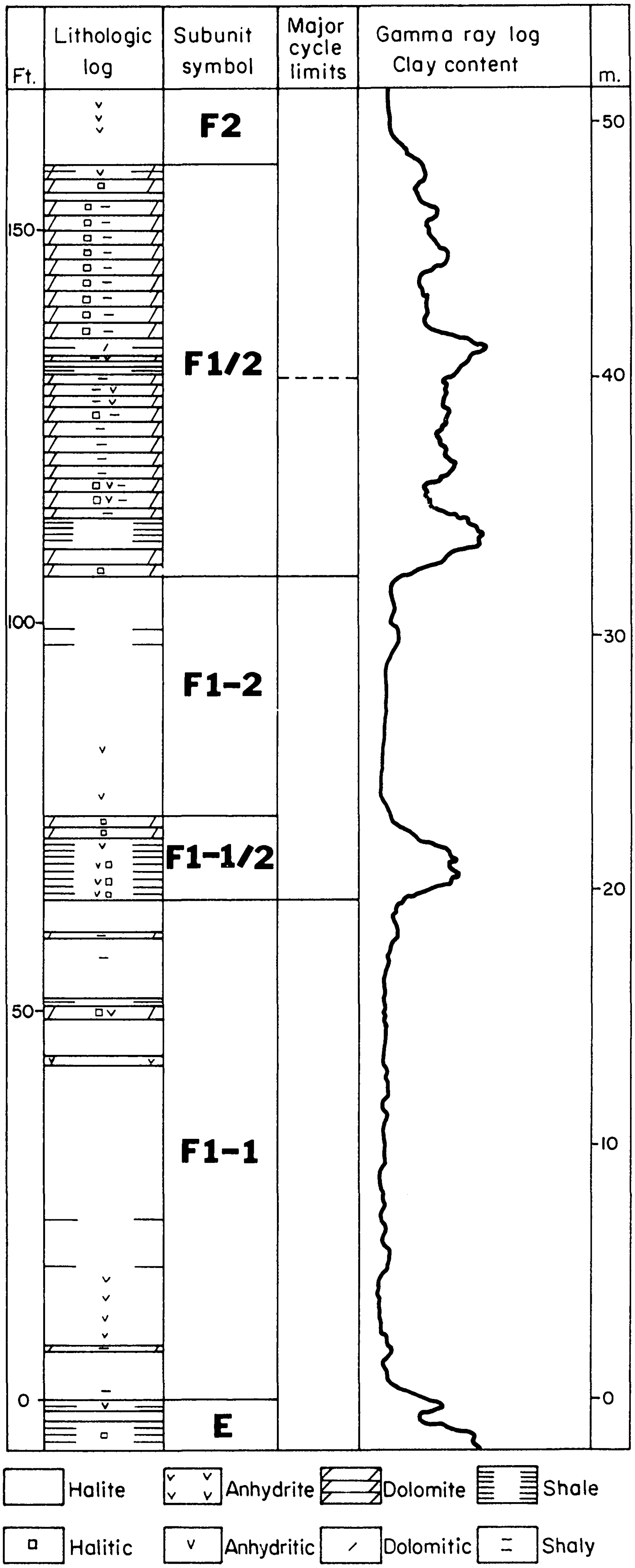


Fig. 49

Upper Syracuse, top section at Watkins Glen, New York

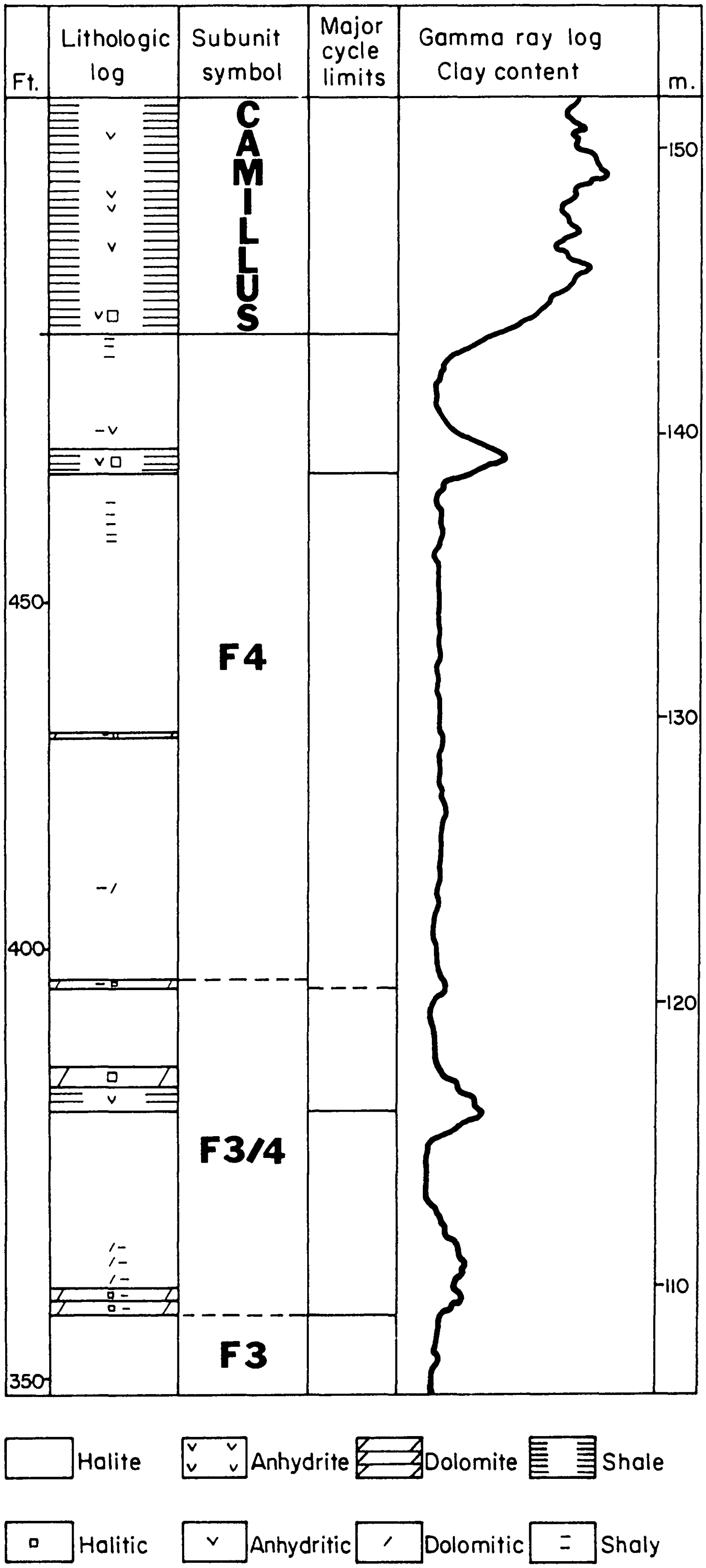


Fig. 50

Figure 51

Ramifying base of dolomite intertwined with the underlying dark salt.

Actual size

Upper Syracuse, Watkins Glen, New York

Figure 52

Light dolomite network in dark salt. Some of the dolomite bodies taper downwards, filling spaces between upward pointing halite crystals, and have relatively flat top surfaces.

Actual size

Upper Syracuse, Watkins Glen, New York

Figure 53

Cavity system in massive dolomite, filled with salt

Actual size

Upper Syracuse, Watkins Glen, New York

Fig. 51

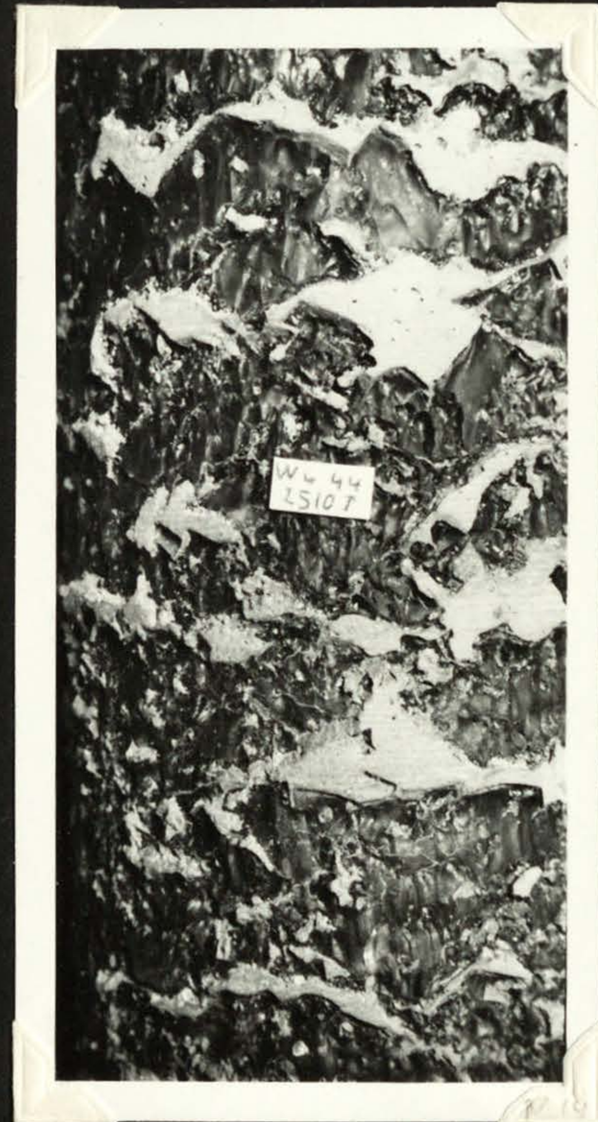


Fig. 52

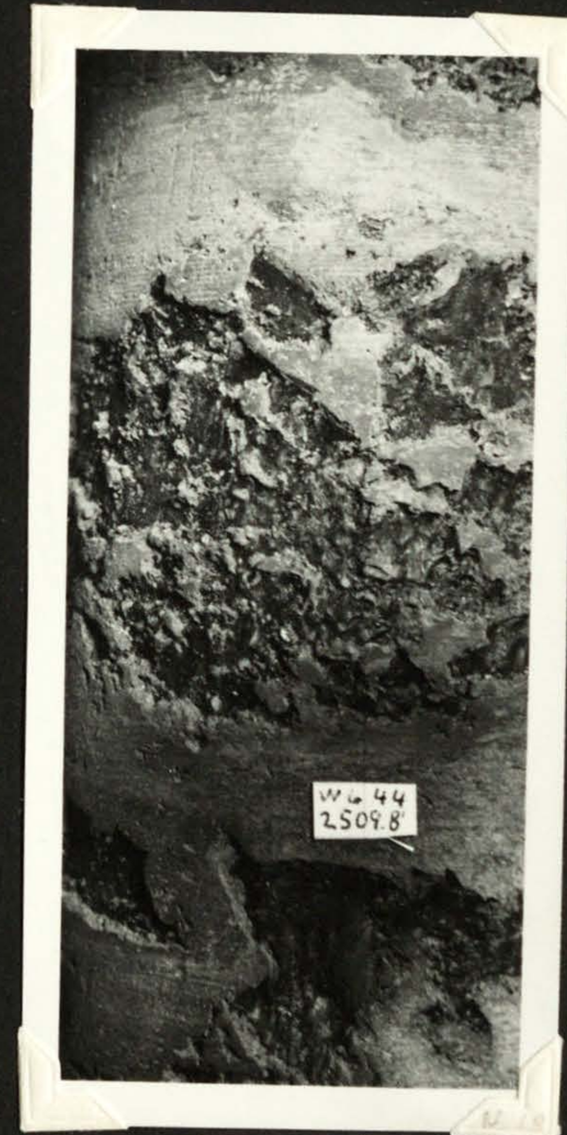


Fig. 53

Syracuse Dolomite Suite. As there, a few quartz grains of slightly smaller size (25-30, seldom 80 microns) also occur. The dolomite contains dark laminae of organic matter, frequent small lenses of halite, and a few anhydrite needles (0.2 mm). Large halite nodules, (10 cm; Figure 53) are interwoven with dolomite network.

Within the salt that overlies the dolomite interbed, shale relics as well as clear crystals and lenses are fairly common; the groundmass consists of normally medium to dark grey halite. Upon hitting, this salt often emanates H_2S. Relatively short uncontaminated intervals are characterized by larger crystals and the lack of shale fragments; in these clear salt beds and lenses, the halite crystals are often lined with a thin film of anhydrite. The contacts of clearer, possibly recrystallized sections and the normal dark halite are often sharp (Figure 54). In the major part of the Fl-1 salt layer, non-halitic interbeds (dark grey shale and argillaceous dolomite) occur rarely; within a thickness of 63 ft, only six such interbeds have been observed. Their thickness ranges from 0.15 to 2.75 ft; four of them are 0.4-0.8 ft thick. These interbeds are characterized by flat, horizontal lamination, few salt lenses, and thin lenses and laminae of anhydrite.

Where the amout of shale or shaly dolomite is high but not sufficient to form an independent bed in the salt, it forms irregular, more or less discontinuous network (Figure 55). Transitions from a dolomite interbed to such a network are characterized by progressively thicker halite nodules, frequently round at their bottom and flat at their top (Figure 56). Minor flowage is often evident.

<u>Figure 54</u>

Clear salt lamina with intercrystalline white anhydrite film

Actual size

Upper Syracuse, Watkins Glen, New York

<u>Figure 55</u>

Shaly dolomite network (matte) in dark salt (bright). The salt surface is slaightly etched

Actual size

Upper Syracuse, Watkins Glen, New York

<u>Figure 56</u>

Dolomite overlying dark salt and containing large nodules and lenses of salt in its basal section.

Actual size

Upper Syracuse, Watkins Glen, New York

Fig. 54

Fig. 55

Fig. 56

Anhydrite occurs only as a subordinate component. It forms a film among the halite crystals of the thin, clearer salt beds, as well as thin lenses and laminae in the shale and dolomite relics and interbeds contained in the darker salt. As a major component, it occurs only once, forming an anhydrite interbed (1 ft) in the halite; even here, the anhydrite is strongly contaminated with dolomite, which becomes dominant in the upper part of the interbed.

F1-1/2 argillaceous dolomite

A major argillaceous-dolomite layer, which separates the lower and upper part of the F1 salt unit, is very persistent; its thickness, only ~10 ft, hardly changes for several miles.

The contact of this argillaceous dolomite layer with the underlying salt is a conspicuously flat, even surface (Figure 57) coated with fine pyrite. The argillaceous dolomite is often well laminated; laminae of shale and of clay-poor dolomite alternate. The rock contains frequent thin lenses and laminae of anhydrite, particularly in the top portion of the layer; thin fissures (<1 mm) filled with anhydrite are also common. Halite lenses are very frequent throughout; occasionally they are connected by minor, channel-like fissures.

The top of the F1-1/2 layer is again bioconstructed: irregularly curving rootlets and undulating laminae consist alternately of dark grey and brown dolomite. Cavities are common between the rootlets, they are filled by salt (Figure 58).

Figure 57

Dark salt overlain by dolomite along comparatively flat contact. The salt contains few clear crystals. The dolomite contains dark argillaceous flakes and white anhydrite lenses, and it is penetrated by a thin network of cracks filled with white anhydrite and recrystallized dolomite

Actual size

Upper Syracuse, Watkins Glen, New York

Figure 58

Laminated dolomite reef rock built chiefly by algae, consisting of irregularly alternating dark and light laminae. Cavities are filled with dark halite.

Actual size

Upper Syracuse, Watkins Glen, New York

Figure 59

White anhydrite bed in dark salt. The anhydrite contains thin argillaceous (grey) laminae containing white anhydrite lenses, as well as grey shale flakes.

Actual size

Upper Syracuse, Watkins Glen, New York

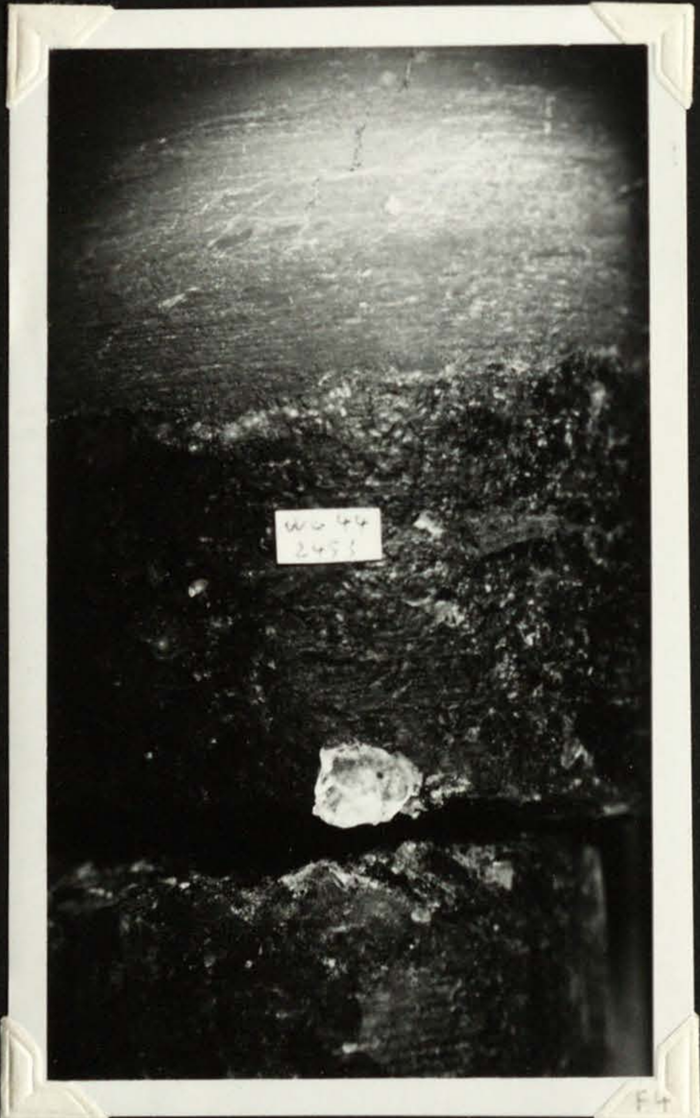


Fig. 57

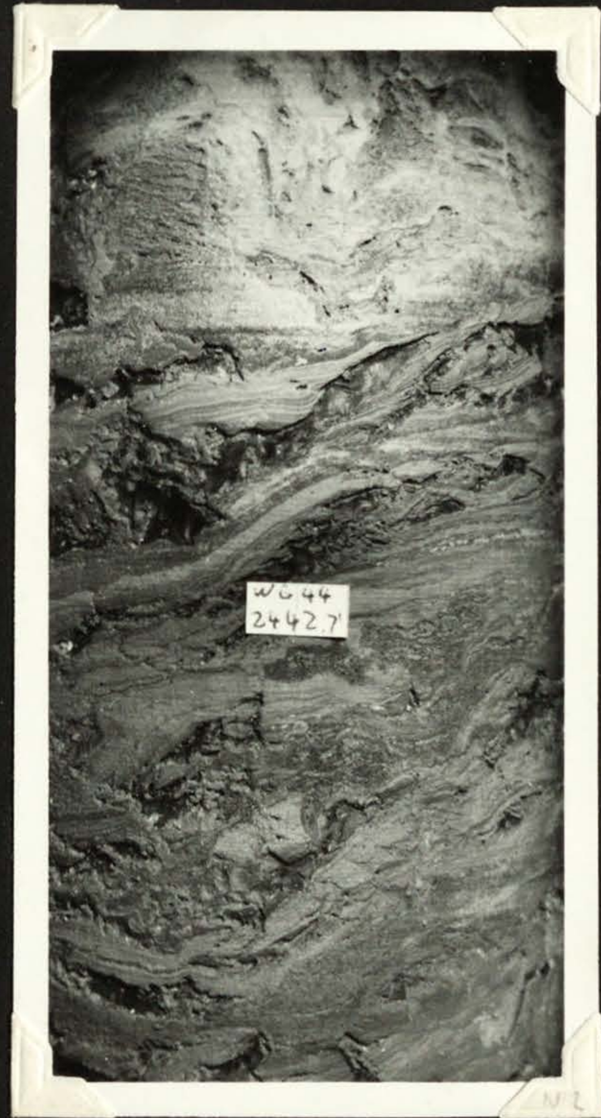


Fig. 58

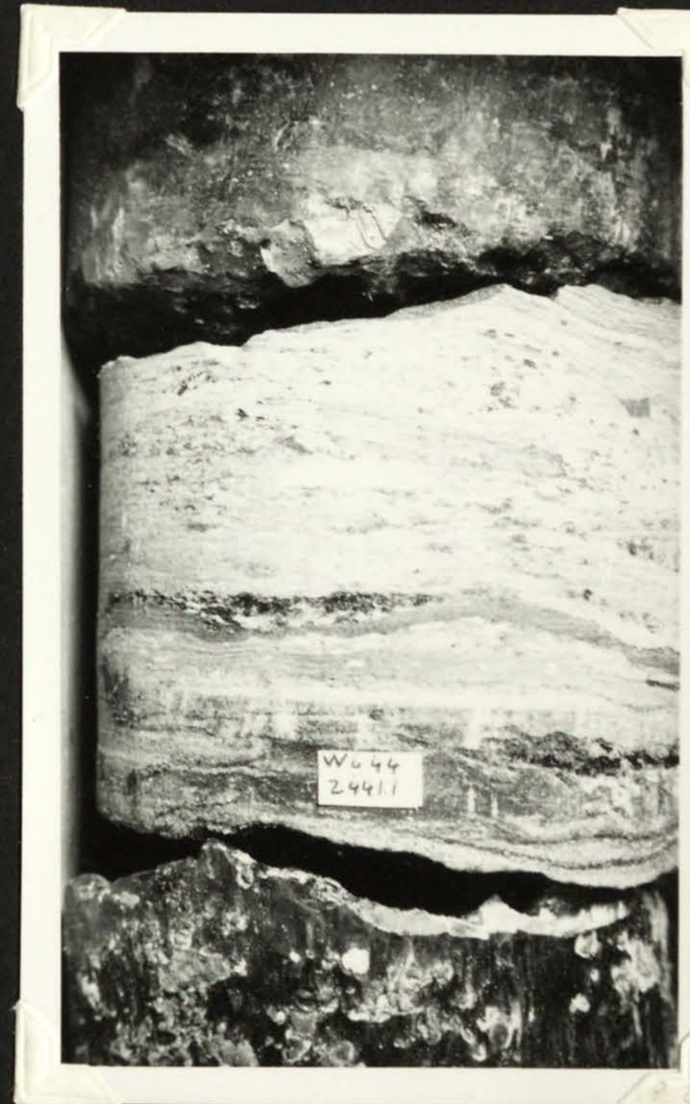


Fig. 59

F1-2 salt

The reef-rock at the top of the F1/2 layer is overlain by salt along an irregular contact. In this relatively coarse (10 mm) light grey salt, which contains no or almost no dolomite relics, a bed and laminae of anhydrite intercalate. Near bottom an anhydrite bed (0.3 ft, Figure 59) overlies the very coarse, lowermost salt bed with a sharp contact and contains severely fragmented laminae of shale and shaly dolomite. The anhydrite itself consists of more or less congruous lenses and laminae, separated by darker, slightly argillaceous-dolomitic intervals.

Further up, but still in the lower portion of the F1-2 halite layer, only a few anhydrite laminae occur. The thickness of these composite laminae is 1-2 cm (Figure 60) , they consist of almost parallel horizontal to slightly undulating anhydrite laminae, 2-5 mm thick, alternating with wedging out halite laminae of similar thickness. The halite layers between these composite laminae range in colour from almost clear to dark grey. Relics of shale and dolomite are still virtually absent in them. In two cases the light halite layers overlying the anhydrite laminae contain intercrystalline anhydrite film (Figure 60,62).

The upper half (about 20 ft) of the F1-2 salt is different from its relatively pure lower section. While the latter contains only some an-

Figure 60

Set of alternating anhydrite (white) and salt (light grey) laminae in coarsely crystalline salt. The thin halite laminae frequently wedge out. In the enclosing coarse halite, anhydrite relics and anhydrite film among crystals are common

Actual size

Upper Syracuse, Watkins Glen, New York

Figures 61, 62

Coarsely crystalline salt, light grey to clear, containing white intercrystalline anhydrite film

Actual size

Upper Syracuse, Watkins Glen, New York

Fig. 60

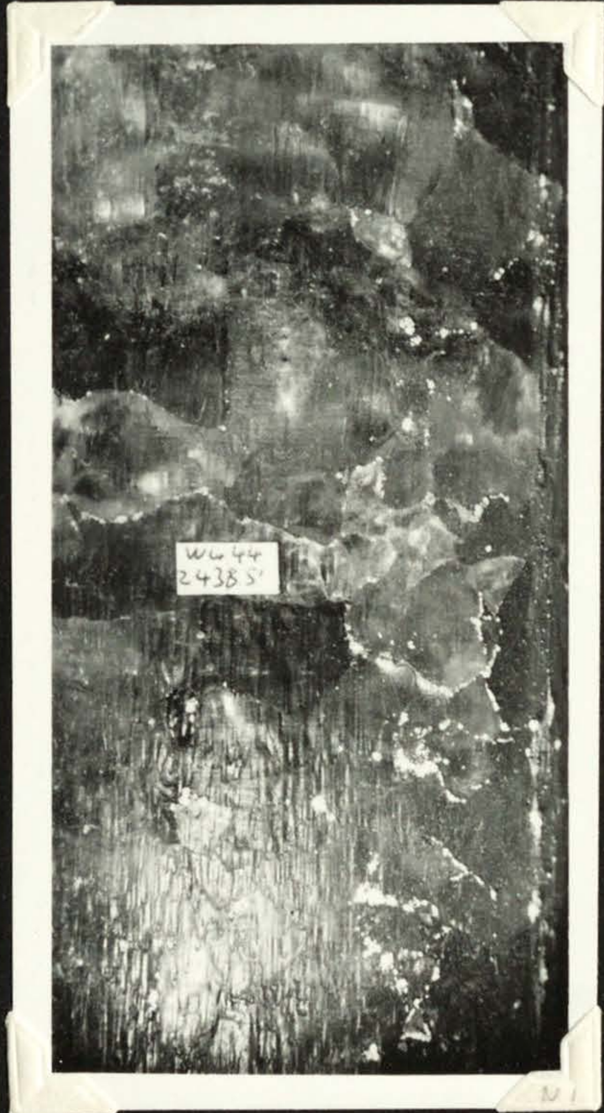


Fig. 61

Fig. 62

hydrite, in the upper portion dolomite and shale are the most common contaminants. They range from isolated shale and dolomite relics in the salt (Figure 63) to continuous shale network with large isometrical salt nodules (Figure 64). Rarely, thin beds of shale and argillaceous dolomite intercalate (Figures 65,66,67) marking a local shift in the shale network - halite nodules ratio. Occasionally, these shale beds show fine internal lamination, due to the oscillation of the dolomite content. Near the top of the F1-2 layer, these shale beds increase in thickness, but even the uppermost is less than 1 ft thick, and is succeeded by more than 4 ft of salt before being overlain by the F1/2 dolomites.

Figure 63

Discontinuous dolomite network in medium grey salt

Actual size

Upper Syracuse, Watkins Glen, New York

Figure 64

Matte dolomite network surrounding bright isometric salt nodules

At bottom right, a fracture cutting through salt nodules and network alike is filled with clear salt.

Actual size

Upper Syracuse, Watkins Glen, New York

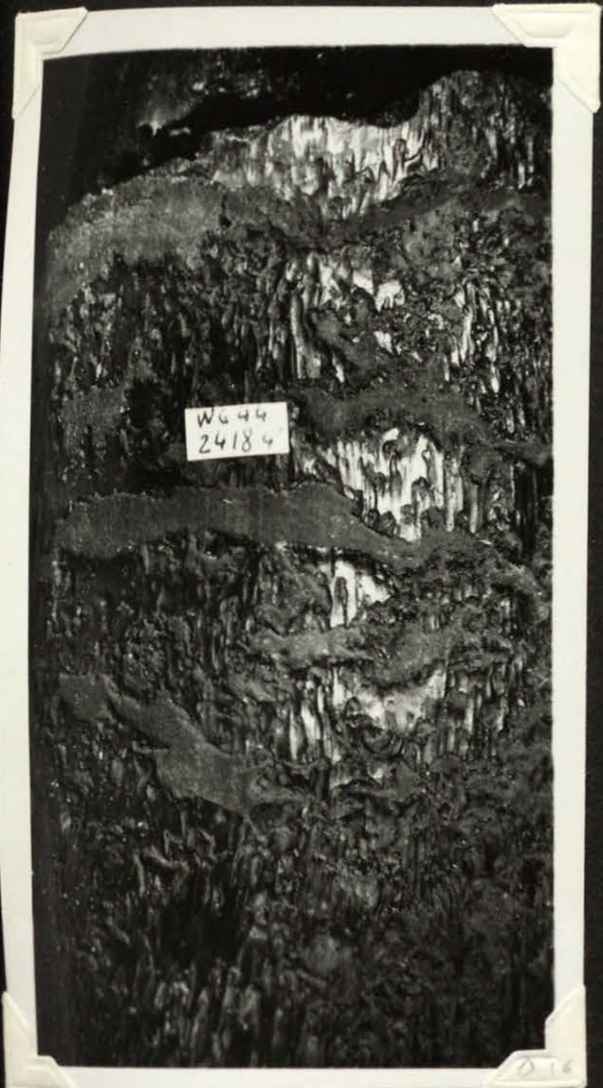


Fig. 63

Fig. 64

<u>Figure 65</u>

Pyramids and cubes of halite (H), resting on argillaceous dolomite bed (D_S) and covered by thick dolomite network (D)

Actual size

Upper Syracuse, Watkins Glen, New York

<u>Figure 66, 67</u>

Dolomite network in salt grading upwards into a laminated dolomite interbed of alternating lighter and darker laminae

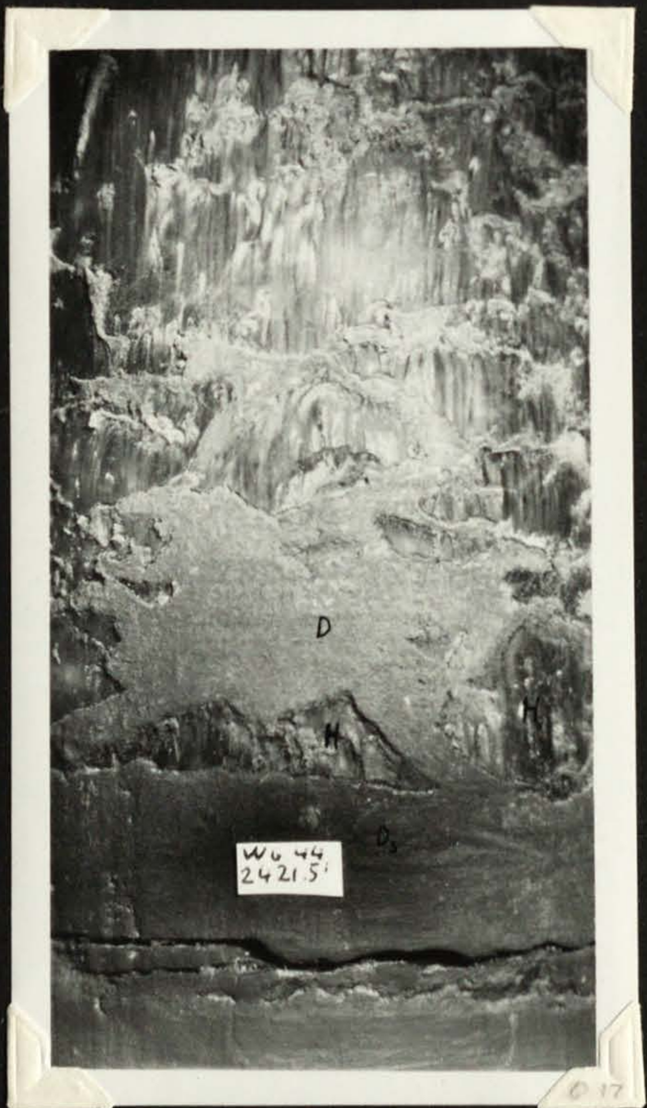


Fig. 65

Fig. 66

Fig. 67

F1/2 dolomite

The F1 and F2 salt units are separated by a sequence of dolomites, more than 50 ft thick. The lower half of this dolomite is strongly argillaceous; it contains shale beds both at its top and bottom. The upper one is less argillaceous and is separated from the lower portion by a series of thin salt interbeds.

Lower, argillaceous section

The dolomite overlies the F1 salt with a very uneven contact. At its bottom, without losing its congruity, the dolomite becomes more and more penetrated by salt protrusions, reaching upwards from, and filled by, the underlying salt (Figure 68), or, looking at it from another angle, the dolomite fills complex-shaped pockets within the halite. Salt lenses, both vertical and horizontal, gradually decrease upwards, while the clay content increases. The rock soon assumes a flat laminar-lenticular structure, though in local, brecciated intervals (Figure 71), irregular lenses and fragments of less argillaceous dolomite are embedded in the more argillaceous matrix. The shale gradually becomes dominant, and thin (-1mm) flat dolomitic shale and argillaceous dolomite laminae alternate, occasionally containing a somewhat thicker (3mm) lamina of halite. The bedding planes often display a polygonal network of thin, halite-filled fissures.

Above this shaly section, clay content again declines, giving place to a new biostromal layer. Irregular, upwards convex, very thin laminae of alternately dark and light dolomite (Figure 69) become in-

<u>Figure 68</u>

Dolomite containing abundant dark halite lenses along an irregular salt/dolomite contact

Actual size

Upper Syracuse, Watkins Glen, New York

<u>Figure 69, 70</u>

Finely laminated algal stromatolite, forming upwards convex "cauliflower" structure overlain by horizontal lamination. On Fig. 68, the stromatolite is dissected by a steep fissure filled with coarse clear halite

Actual size

Upper Syracuse, Watkins Glen, New York

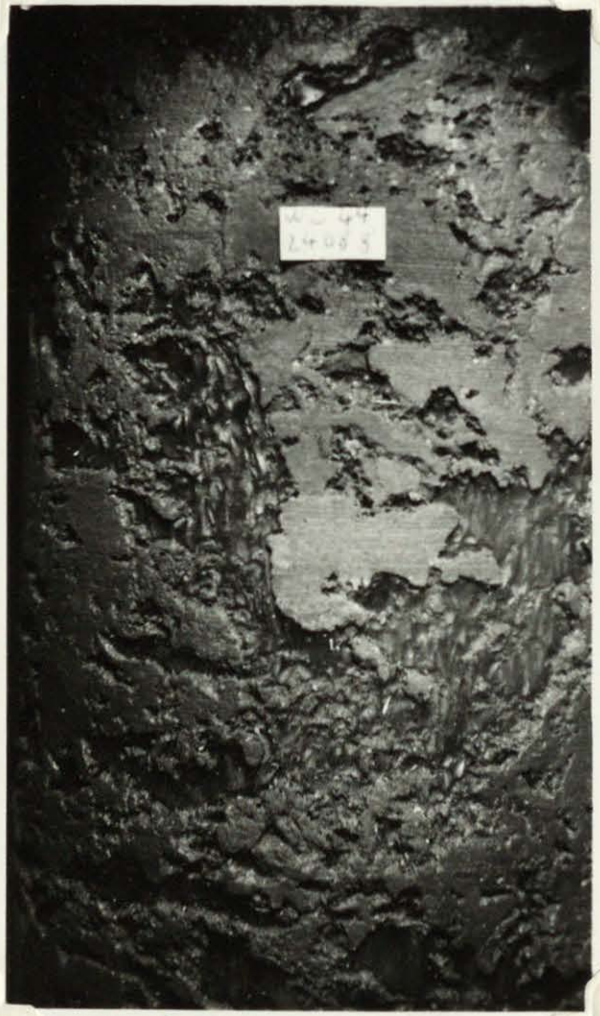


Fig. 68

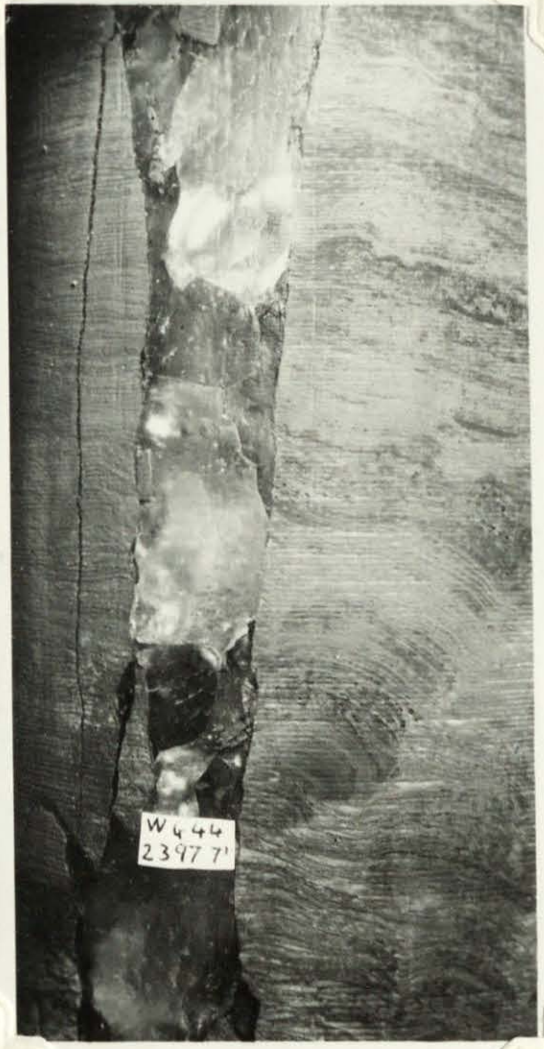


Fig. 69

Fig. 70

creasingly convex at top, with strongly contorted lamina-sets. These "folded" sets are covered by undisturbed laminae (Figure 70). Occasionally the thin laminae are alternately rich and poor in halite and form single mounds in almost massive salt-free dolomite.

Gradually the biostromal structures disappear, while the clay and anhydrite content increases. Distinct lamination assumes a different aspect: generally flat laminae of dolomitic shale and shaly anhydrite alternate, with white anhydrite lenses intervening. The flat anhydrite lenses consist of smaller lenses separated by shale films (Figure 72). Though thin fissures are generally filled with salt veins, shallow crevices filled with anhydrite occasionally also appear. Pyrite is a common accessory mineral, it occurs both in the argillaceous anhydrite and in the shaly partings.

Some laminae show slight brecciation: under the microscope slightly rounded flat flakes of dolomitic shale and of halite are embedded in the dolomite. Some of the flakes consist of slightly laminated shale with large recrystallized dolomite rhombohedra, others of halite or halite and dolomite, also sometimes with recrystallized dolomite rhombohedra. Concentration of pyrite along the margin of the flakes is frequent. (Fig. 104-105).

Gradually, fine lamination disappears, only shaly partings separate the thick dolomite beds. Clay content suddenly rises again and a massive grey shale concludes the lower, argillaceous portion of the Fl/2 dolomite.

Figure 71

Intraformational dolomite breccia in darker, argillaceous matrix in the basal, argillaceous section of a dolomite suite (F1/2)

Actual size

Upper Syracuse, Watkins Glen, New York

Figure 72

Anhydritic transitional zone between biostromal dolomite and overlying argillaceous beds. Laminae of dolomitic shale and argillaceous dolomite (grey) alternate with laminae and lenses of anhydrite (white). Shallow crevices and depressions filled with the material of the overlying lamina are common

Actual size

Upper Syracuse, Watkins Glen, New York

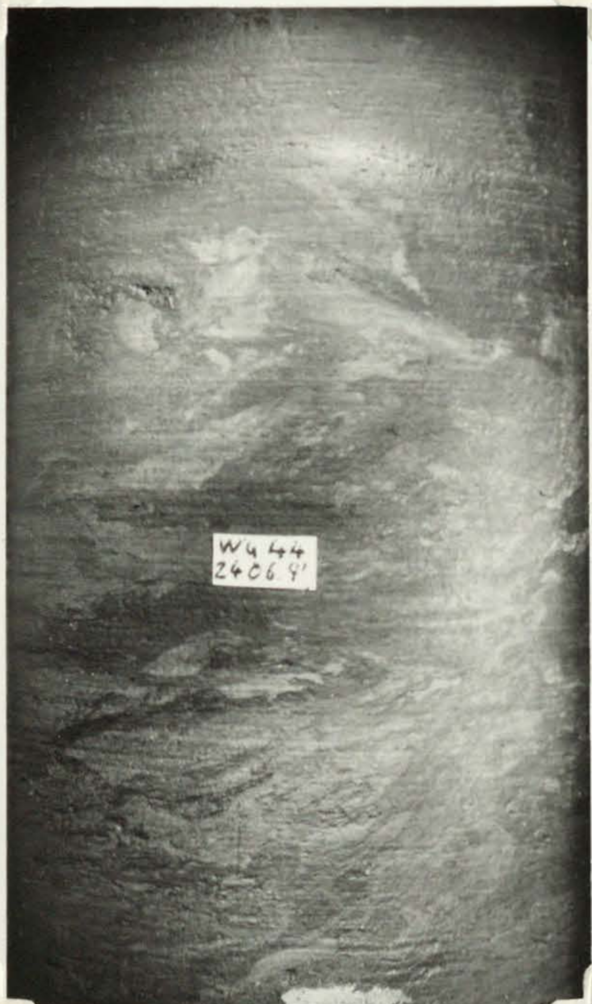


Fig. 71

Fig. 72

<u>Figure 73</u>

Argillaceous dolomite interbed with indistinct lamination in halite suite

Actual size

Upper Syracuse, Watkins Glen, New York

<u>Figure 74</u>

Bottom part: Dolomite containing white aggregates of small anhydrite lenses and dark argillaceous wisps

Top portion: Intraformational dolomite pebbles embedded in darker, argillaceous dolomite

Actual size

Upper Syracuse, Watkins Glen, New York

<u>Figure 75</u>

Light dolomite nodules of probably algal origin, surrounded by branching dark argillaceous laminae. Cavities are filled with dark salt.

Actual size

Upper Syracuse, Watkins Glen, New York

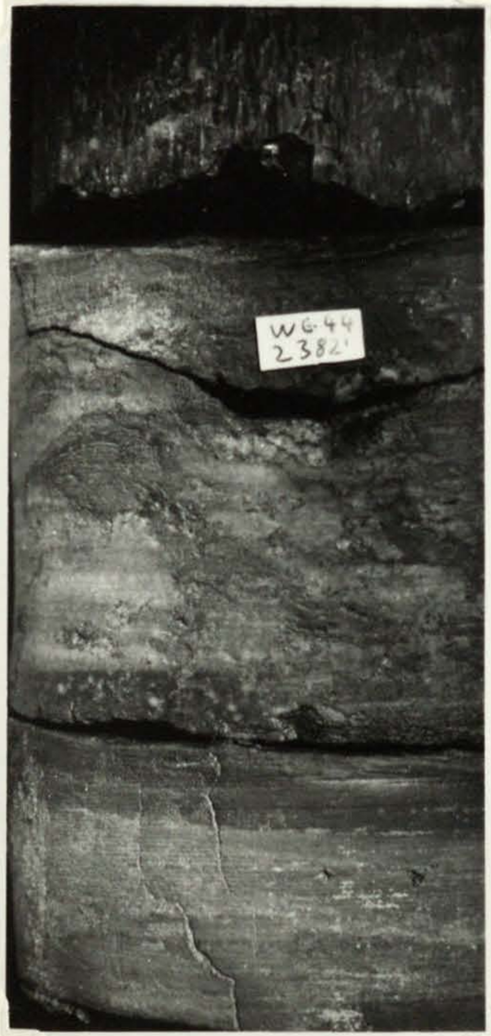


Fig. 73

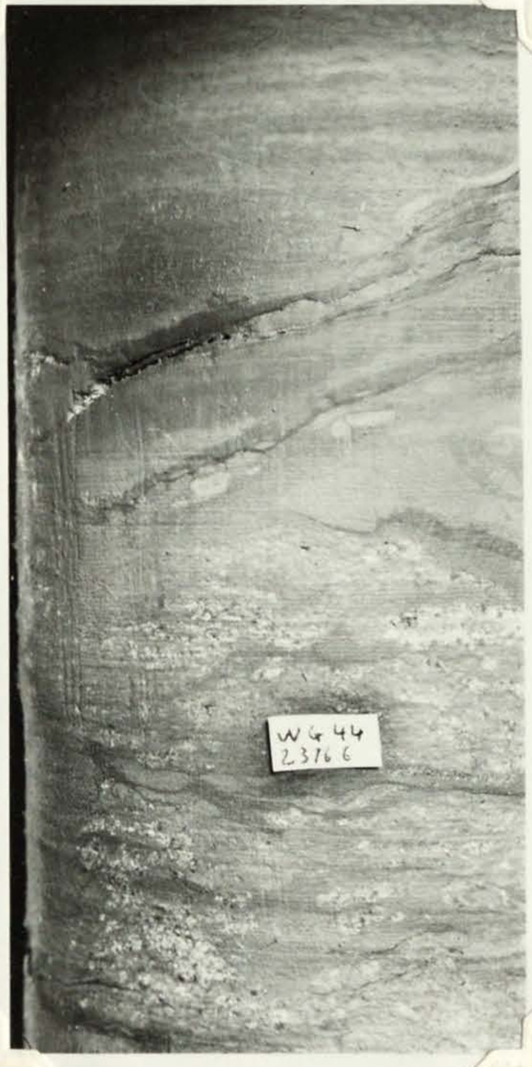


Fig. 74

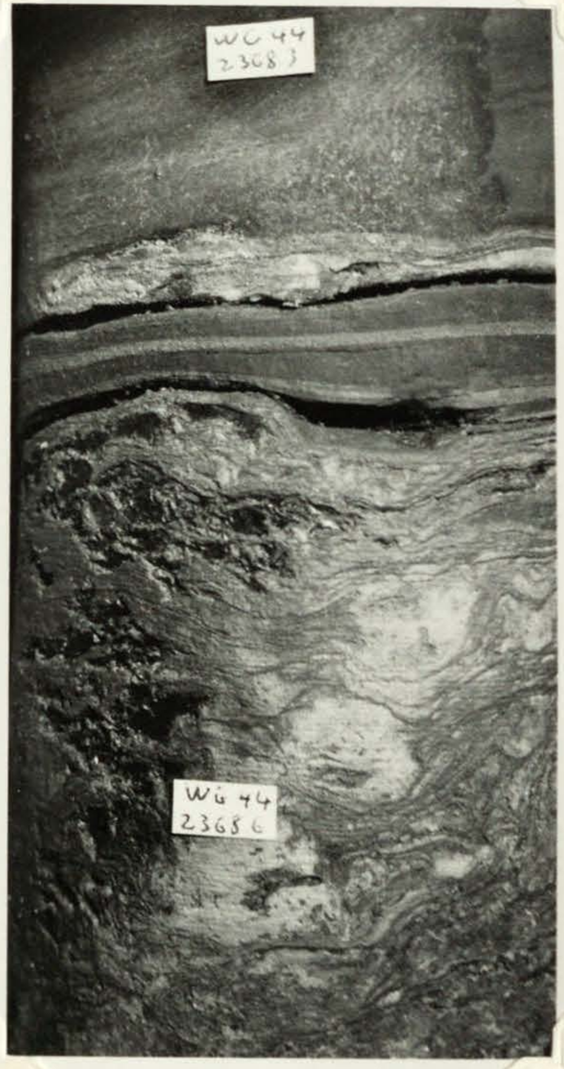


Fig. 75

Salt interbeds

The top shales of the lower, more argillaceous portion of the F1/2 dolomite are separated from the upper, less argillaceous part by a series of three relatively thin salt beds (0.8-1.1 ft), separated by intercalating argillaceous dolomite beds. The salt is light grey, relatively coarsely crystalline in the two lower salt beds and somewhat darker in the uppermost, third one. It contains small relics of shale and argillaceous dolomite in small amounts, mostly in the uppermost, darker bed. The intervening argillaceous dolomite beds (Figure 73) are still thinner (0.3 ft); their oscillating clay content gives rise to moderately distinct lamination. Lenses and fragments of anhydrite are common in them.

Upper, clay-poor section

With a very flat contact, the upper salt interbed is overlain by the upper dolomite. First irregular lenses of dolomite are separated by wavy argillaceous bands; anhydrite lenses, containing euhedral anhydrite porphyroblasts, are common.(Figure 74). Gradually, biostromal structures develop: first the laminae become irregular, undulating, enclosing round dolomite structures; irregular cavities are filled with halite (Figure 75). Further up, the anastomosing laminae become less irregular (Figure 76, 77). Ultimately, the fine flat lamination returns, the laminae, -1 mm, differ in organic and halite content. Thin beds of irregular lamination, with salt-filled cavities (Figure 78) occasionally recur.

<u>Figure 76-77</u>

Undulating dark laminae above a zone of probably algal nodules in slightly argillaceous dolomite

Actual size

Upper Syracuse, Watkins Glen, New York

<u>Figure 78</u>

Bands of alternately horizontal and irregular lamination in finely laminated algal stromatolite. Cavities are filled with dark salt.

Actual size

Upper Syracuse, Watkins Glen, New York

Under the microscope, thicker laminae of organic matter and clay-free dolomite alternate with thinner organic-rich laminae. The rock contains interstitial lenses of halite, and some dispersed pyrite. The thin, more carbonaceous laminae occasionally form thin, flat, wavy networks enclosing purer dolomite lenses. The dolomite of the proportionally more significant organic matter- and clay-free laminae varies in grain size: laminae of coarse-grained dolomite silt (0.03-0.05 mm), like those occurring in and directly above the Middle Syracuse, alternate with the dominant fine-grained dolomite (3-10 microns). The coarse dolomite laminae contain a few quartz grains of similar size. The coarser laminae overlie the fine ones along a sharp, somewhat uneven contact, and pass gradually into the overlying fine laminae which contain only a few coarser grains.

Separated by recrystallized salt (0.7-0.2 ft) filling interstratal fissures, the uppermost part (-2ft) of the dolomite is horizontally bedded, essentially poor in organic matter and clay. It is penetrated through and through by thin, short (1-2 cm) fissures filled with anhydrite. On the bedding plane, these fissures form a polygonal pattern. Halite crystals are relatively common (Figure 79,80).

The F1/2 dolomite contacts the overlying F2 salt with an anhydritic boundary zone. Irregular argillaceous dolomite fragments (Figure 81) are embedded in anhydrite matrix; salt lenses and single crystals in it are frequently connected by thin, salt-filled fissures. Above this anhydrite, a light, coarse halite bed (0.7 ft) contains frequent relics of dolomite; it is separated from the F2 salt by a thin bed in

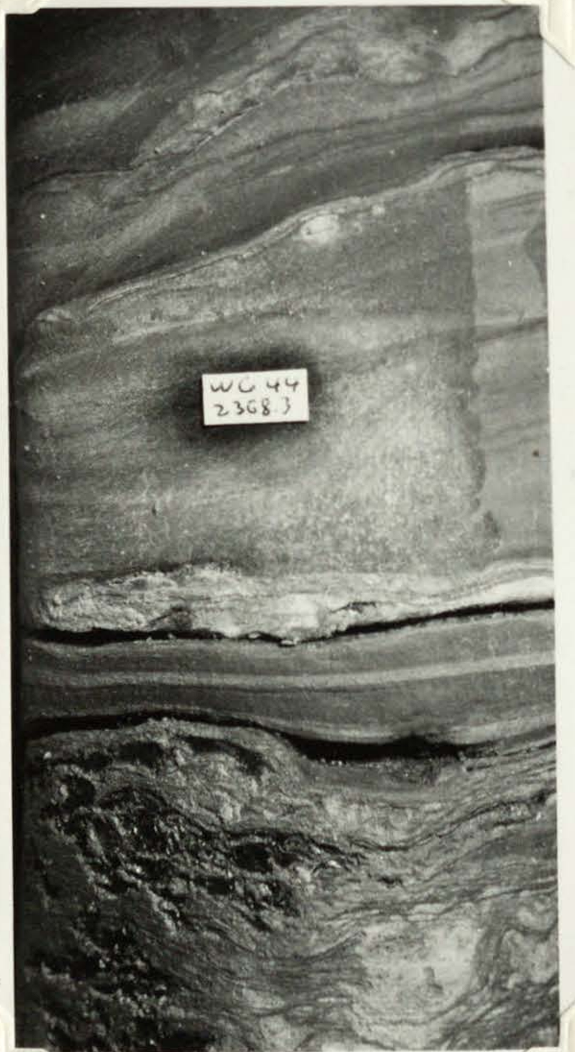


Fig. 76

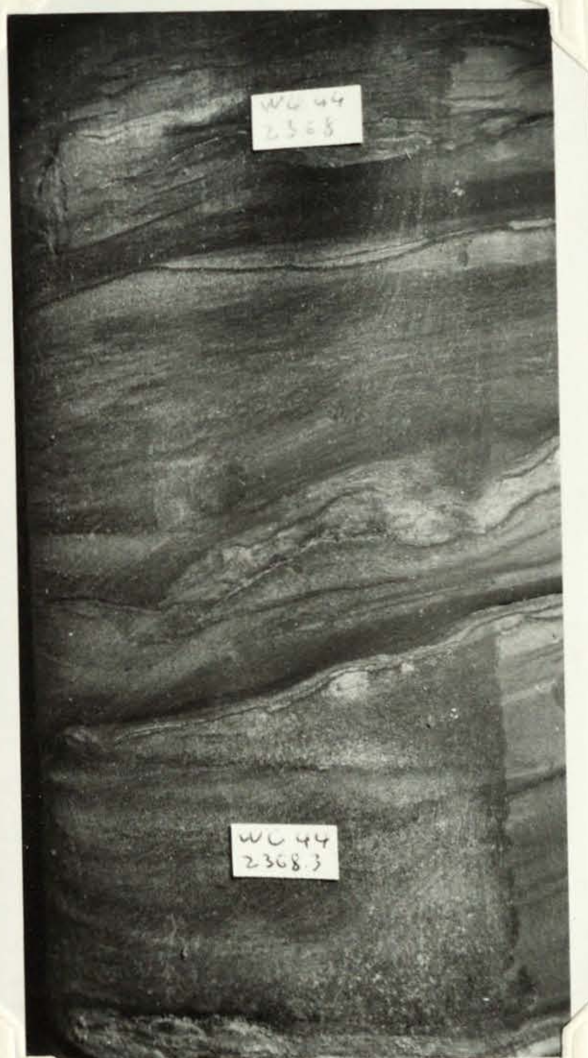


Fig. 77

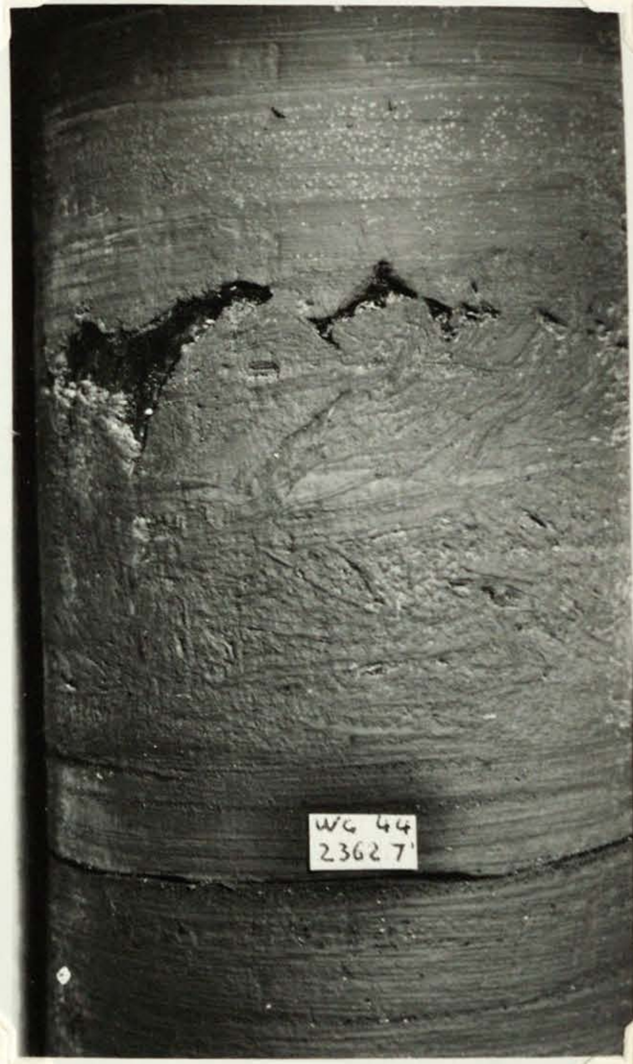


Fig. 78

Figures 79-80 (80 is the upward continuation of 79)

Dolomite with upward increasingly frequent paper-thin anhydrite laminae (light grey) and with intricate crack system filled with white anhydrite. Together with the latter, the rock is slightly fragmented at bottom; these later cracks are filled with halite. At bottom left, wide crevice is filled with clear salt.

Actual size

Upper Syracuse, Watkins Glen, New York

Fig. 79

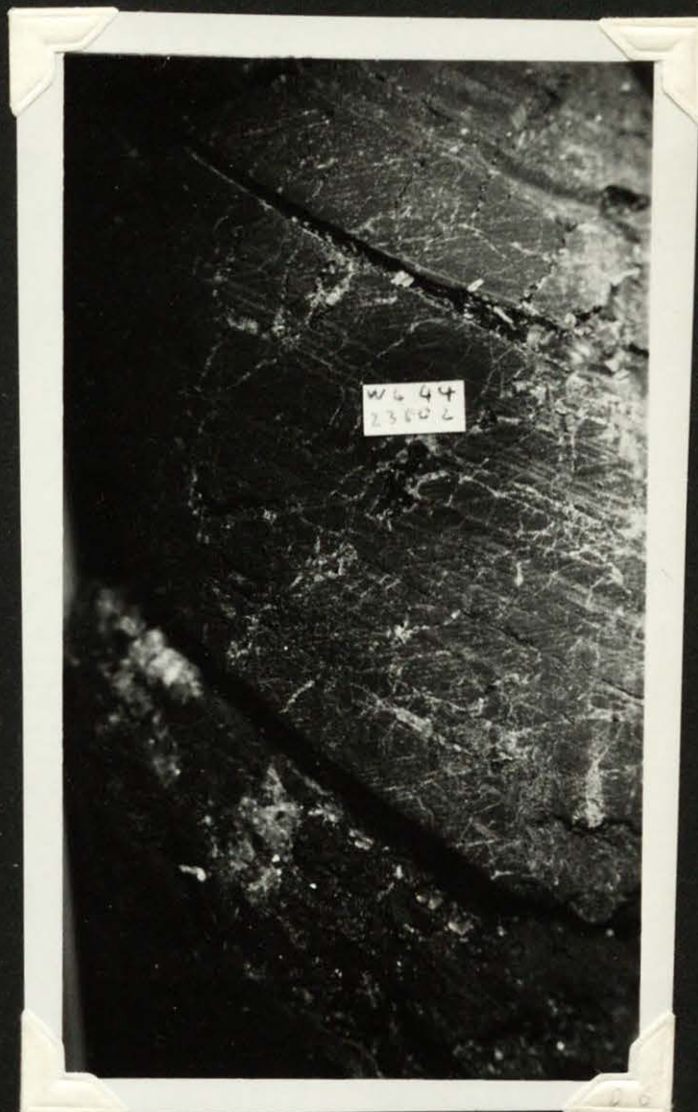


Fig. 80

which shale and dolomite fragments sit in an anhydritic matrix, together with which they are slightly re-brecciated in the salt (Figure 82). These fragments are penetrated by, contain lenses of, and are embedded in grey salt.

Figures 81, 82

Anhydritic boundary zones between halite and overlying argillaceous dolomite, displaying angular fragments of dark argillaceous dolomite embedded in white anhydrite matrix. On the right picture, this breccia is slightly re-brecciated, the resulting crevices being filled with dark salt

Fig. 81

Fig. 82

F2 halite

The halite overlying the F1/2 dolomites is dark grey, and contains varying amounts of clear coarser crystals, singly as well as in lenses and laminae (Figure 83). Temporarily these crystals become dominant, forming a bed of coarse light salt (Figure 84) whose crystals are coated with anhydrite film. Only after renewed decrease in the amount of clear crystals, when the halite turned dark again, do the first shale and argillaceous dolomite relics appear (almost 20 ft above base). Together with these relics, randomly oriented fragments of a thin (6 cm) argillaceous dolomite bed, finely laminated, also occur; they "float in the salt" partially surrounded by clear halite crystals.

Under the microscope, the argillaceous dolomite relics are coated with overgrowing anhydrite and secondary dolomite crystals, contain wavy bituminous shale laminae, and anhydrite lenses occasionally resembling pseudomorphs after gypsum. The anhydrite film among the crystals is also occasionally accompanied by secondary dolomite crystals.

Thin beds and laminae of shale occasionally appear in the salt. These beds are usually somewhat brecciated due to their tectonic incompetence with the salt, but still preserve their congruity. Darker, more argillaceous-bituminous, and lighter, more dolomitic laminae frequently alternate; irregular halite lenses, often perpendicular to lamination, are common, crystals and lenses of anhydrite are frequent. Short discontinuous fractures, possibly desiccation cracks, are filled with anhydrite and halite.

Figure 83

Dark salt with frequent clear crystals

Actual size

Upper Syracuse, Watkins Glen, New York

Figure 84

Band of coarsely crystalline clear salt, in medium crystalline grey salt containing white anhydrite film among the crystals.

Actual size

Upper Syracuse, Watkins Glen, New York

Fig. 83

Fig. 84

In the uppermost 15 ft of the F2 salt, the shale (and argillaceous dolomite) laminae are very halitic. Locally, microscopic laminae of pure and dolomitic halite alternate. Some of these have strongly channelled, sharp, erosional basal contacts, cutting deep into the underlying salt lamina. The concentration of finely dispersed dolomite micrites in the salt rises sharply immediately below each of these erosional surfaces,possibly indicating their enrichment through partial dissolution. Curved bands, with chevron-like internal structure may be traces of algal stemsdragged along the surface by currents. Probable algal stems are sometimes preserved: they lie at a small angle to the lamination and resemble the bifurcating tubes mentioned in the discussion of the Lower Syracuse (D1/2). (Fig. 129-130).

In another thin bed within the salt, only 2 ft below its top, the microscopic laminae consist alternately of relatively pure dolomitic shale and impure halite. The latter laminae, 3-10 mm thick, display intricate, rootlike networks of salt, with dolomite preserved in between the rootlets in the form of similar rootlets, relics, and round grains. Even within the same lamina, dolomite grades horizontally into halite in one thin section. Anastomosing microlaminae of dark organic matter in the dolomitic shale enclose lenses of dolomite and often terminate abruptly. The structure of the rock is frequently pelletal; dark, more argillaceous pellets (30-80 microns) are cemented with lighter dolomite. The shale laminae contain small halite lenses and some quartz silt.

F2/3 dolomite

Subunit F2/3, the uppermost thick dolomite suite of the Syracuse Formation, is generally about 45 ft thick in the Watkins Glen area. Due to a fault, its lower portion is missing from our standard borehole (WG 44), so the discussion of this portion will be based on nearby wells.

The F2 salt is overlain by a thick shale layer (-16 ft); the frequency of large lenses of coarse dark salt gradually decreases upwards. The shale displays fine lamination (with the laminae differing in their organic and dolomite content), and contains pyrite. Above this basal shale, the dolomite content gradually increases. Dolomitic shale and argillaceous dolomites follow, making up about a half of the subunit. Halite becomes increasingly common, filling thin crevices, forming lenses and seldom laminae; anhydrite lenses also occur.

The purest dolomite section is brownish grey. Much of the dolomite is laminated. Laminae containing large amounts of small (-0.3 mm) flattened halite oolites alternate with laminae containing dark dolomite oolites of similar size; the latter often have a lighter core (Fig. 106-108). These oolite-bearing laminae, poor in clay, alternate with argillaceous-anhydritic, but oolite-free, dolomite laminae. Recrystallization is intensive: crystals of secondary quartz and dolomite line, together with halite lenses, the dolomite/shale interfaces, and large secondary dolomite crystals locally envelop entire dolomite laminae.As in other instances of recrystallization, the carbonate and silica have been leached from the dolomite and shale laminae by concentrated interstitial solutions, which ultimately filled the pores with halite.

The upper part of the sequence (about 7 ft) consists again of shale, with frequent laminae and thin beds of argillaceous dolomite. Locally, narrow cracks in the dolomite laminae are filled with the material of the overlying shale lamina. Fragments of dolomite are frequently embedded in the shale, particularly in its lower part. Irregular shale flakes embedded in more dolomitic shale matrix also occur. Lenses and laminae of anhydritic shale are very common, sometimes lenses and isometric concretions of anhydrite occur in their centre. The amount and size of the anhydrite lenses also decreases upwards. Some halite and pyrite crystals occasionally occur; thick fissures, both vertical and interstratal, are filled with clear crystals of halite.

The top 3.5 ft of the F2/3 subunit consists of a halite and above it a dolomite bed, overlying the shale and marking the transition into the F3 salt. The halite is light grey at bottom and becomes medium grey towards its top; it contains shale laminae and relics near its basal and a few dolomite relics near its top contact. In one of the wells, the salt contains a bed of shaly anhydrite, with intensely distorted thin shale laminae. The dolomite is argillaceous, more shaly toward the SE; it contains only a few thin lenses of halite.

<u>Figure 85</u>

Photomicrograph of dark relics of argillaceous dolomite in halite. The extensive lighter area of the largest dolomite relics suffered partial replacement by halite.

Nicols not crossed; magnification 12 diameters

Upper Syracuse, Watkins Glen, New York

<u>Figure 86</u>

Light argillaceous dolomite relics in dark halite

Actual size

Upper Syracuse, Watkins Glen, New York

<u>Figure 87</u>

Large recrystallized clear crystal in dark halite

Actual size

Upper Syracuse, Watkins Glen, New York

Fig. 85

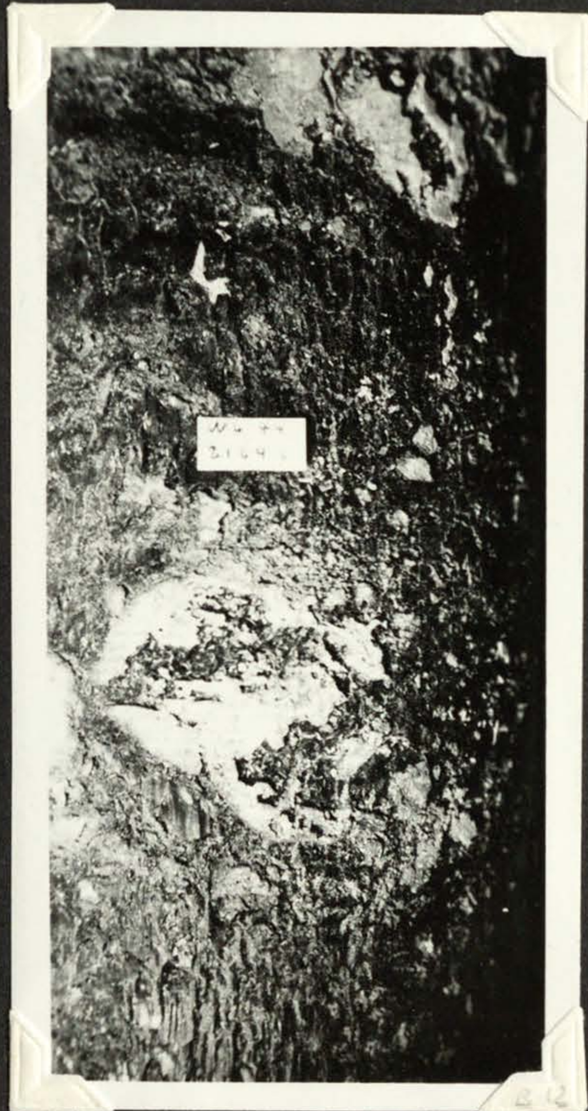


Fig. 86

Fig. 87

F3 and F4 salts

The upper part of the Syracuse formation is a thick suite of halite (normally 165 ft at Watkins Glen). Occasionally, it is separable into two subunits, F3 and F4, but the shale beds between them are thin and difficult to correlate.

The salt overlies the F2/3 dolomite with a relatively flat basal contact. There are only few dolomite relics near the base, but soon they join to form a continuous network. Under the microscope, the network consists of argillaceous dolomite, and contains lenses of anhy - drite. At places the network thins into an intercrystalline film consisting largely of anhydrite. Locally, some of the argillaceous dolomite relics are impregnated and partially replaced by halite (Microphoto 85). Laminae and lenses of recrystallized clear halite are rare. The frequency of argillaceous dolomite relics slightly decreases almost throughout the lowermost halite layer (about 21 ft); only near its top shows it again an abrupt increase. Such cycles of slowly decreasing and rapidly increasing amounts of shale or dolomite relics recur repeatedly in the overlying halite layers.

In general, thick halite layers (10-20 ft) are separated by thin interbeds of dark argillaceous dolomite and dolomitic shale (0.3-0.7 ft). In some of these, straight laminae of dolomite alternate with thinner (mm) laminae of dark shale; thin dolomite laminae with small flat salt-filled cavities display organic structures (stromatolites,etc). Lenses and laminae of anhydrite are common; some of these contain dolomite fragments brecciated in and re-cemented by the anhydrite lenses. Generally, halite lenses are equally common. The fragments of an occasional bryozoa(Hallopora? sp) are separated and partially replaced by halite. Many of these thinner dolomite beds are covered by a shale lamina separating it from the overlying shale layer; the shale contains thin laminae and small lenses of anhydrite. In the salt itself, minor flowage broke up some of the shale relics and interbeds, otherwise they contain few halite crystals.

In the purest 50 ft of the F3 halite, shale interbeds are absent; only temporary dominance of shale relics and networks (Figure 86) divides the halite into thinner "layers". Here the halite is often very dark and in several intervals lacks in both clear crystals and shale relics. Otherwise the limits of recrystallized clear crystals and aggregates are always very sharp (Figure 87); very coarsely crystalline (-5 cm) clear salt often fills fissures in the dark salt beds.

Above this interval, argillaceous dolomite interbeds (0.7-6.0 ft) alternating with halite separate the F3 and F4 subunits. Their base consists of shale overlying the halite at a relatively sharp contact; many of the frequent anhydrite lenses in this shale contain discontinuous shale flakes. Dark flakes are common in the somewhat lighter shale as well, as are lenses and crystals of halite. In the upper half of the beds the shale gives place to dolomite. In some beds this dolomite is brown and contains abundant halite spherules. In other beds, the dolomitedisplays a stromatolitic structure, with lighter and organic-rich darker laminae alternating. Some wedging-out laminae present a cellular structure with flat, salt-filled cavities. Near the top, sometimes the bedsbecome again more argillaceous and contain fragmented shale flakes. They pass gradually into the overlying halite with increasing halite nodules and decreasing argillaceous dolomite relics between them.

Above these interbeds, a thick section of relatively pure salt follows (F4, 70 ft). Here even thin interbeds of dolomitic shale are rare. The bottom interval, overlying the upper thick dolomite interbed, is particularly clear, and relatively coarsely crystalline. A few feet above its base, two laminae of anhydrite occur, respectively 0.5 and 3.0 cm thick. The anhydrite contains fragments of horizontally lying to slightly bent shale laminae. The upper anhydrite lamina is "corroded" at its bottom and intruded by halite crystals (Figures 89-90).

Further up, the normal argillaceous dolomite network displaying negative salt crystal forms and containing small lenses of anhydrite (Figure 88) becomes common again, leading into one of the thin dolomite beds of the F4 salt. A peculiar feature of this argillaceous dolomite bed, which contains a few dark halite lenses, is the presence of an upward tapering inverted pocket at its bottom, filled with clear halite and joining the underlying dark salt (Figure 91).

The F4 salt is characterized by oscillations in the amount of dolomitic shale relics. Lenses, laminae and crystals of clear recrystallized halite are rare. The shale occasionally joins into continuous network; a few thin argillaceous dolomite beds with lenses of halite also occur. Here alone does the increase in the amount of shale relics proceed upwards at the same low rate as their gradual decrease.

Close to its top, thin beds of lighter halite occur occasionally in the salt; the amount of dolomitic shale relics is frequently very

Figure 88

Intercrystalline matte shale network, containing small white anhydrite lenses, in dark halite. The network emphasizes the cubic shape of the crystals.

Actual size

Upper Syracuse, Watkins Glen, New York

Figure 89

White anhydrite (A) lamina, enclosing thin laminae, lenses and flakes of dark shale (S) in grey halite (H).

Upper Syracuse, Watkins Glen, New York

Figure 90

Set of white anhydrite (A) and grey halite (H) laminae in dark salt. At the bottom of the laminaset, the anhydrite laminae are thin and discontinuous; upward they become thicker and continuous, possibly indicating a trend of pulsatingly decreasing brine concentration. The enclosing salt contains large clear crystals of recrystallized halite (H_2).

Actual size

Fig. 88

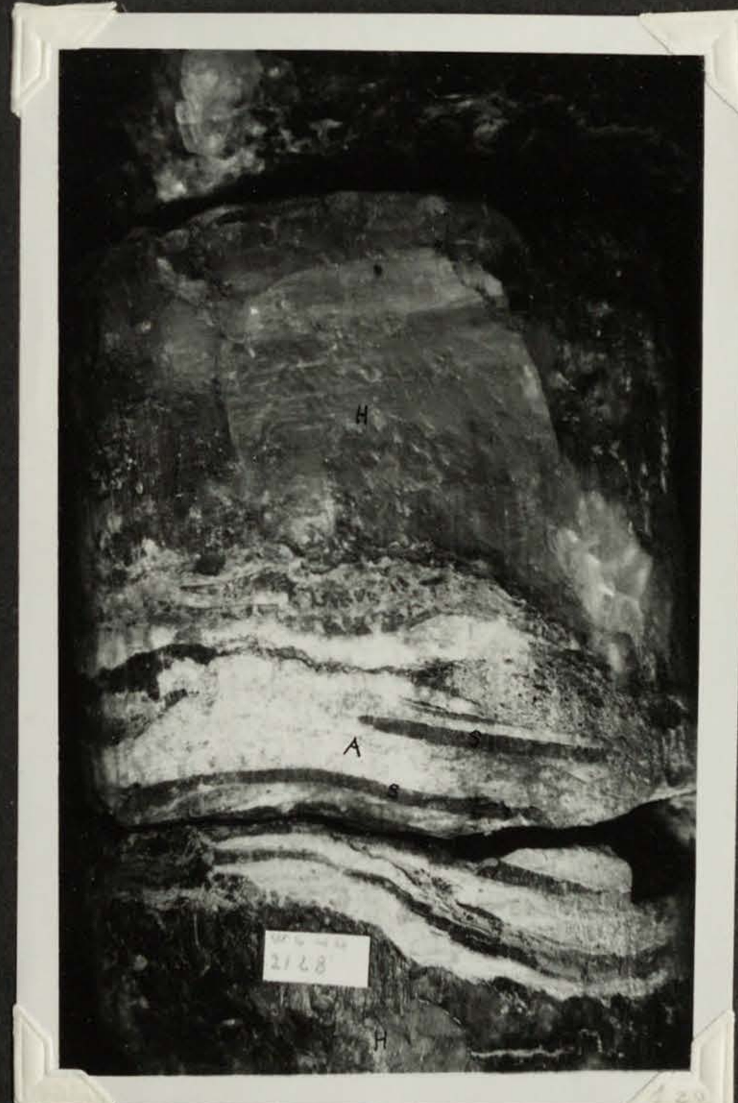


Fig. 89

Fig. 90

Figure 91

Tongue of recrystallized clear salt (H_2) intruding from the underlying dark salt (H) layer into dark argillaceous dolomite (D_S) containing dark halite (H) lenses.

Actual size

Upper Syracuse, Watkins Glen, New York

Figure 92

Band of clear recrystallized salt (H_2) in dark salt (H) containing anhydritic shale (S_A) network

Actual size

Upper Syracuse, Watkins Glen, New York

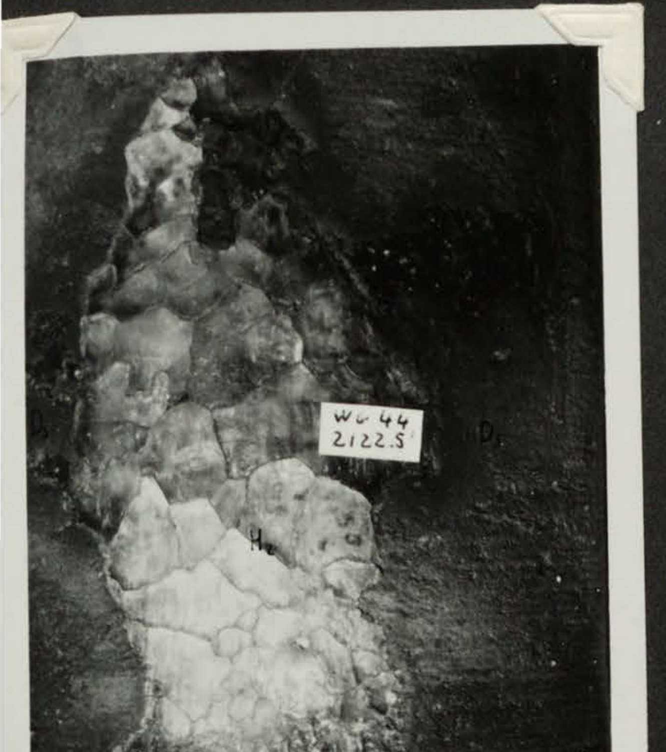


Fig. 91

Fig. 92

low. Locally, bands of clear recrystallized salt cut across the dark salt; they are free of the shale relics and networks contained in the surrounding salt (Figure 92).

Close to the top of the Syracuse formation, a thicker shale bed (2 ft) occurs. It contains a zone of large anhydrite nodules (Figure 93). While in the shale, below this zone, mainly halite lenses and nodules occur, above it undulating laminae and lenses of anhydrite dominate over halite lenses cutting across both the anhydrite and the shale.

Above this uppermost shale interbed, halite dominates once again, though shale relics are still common. Though this uppermost salt layer (about 9 ft) contains a thin clear bed with anhydrite film among its crystals, the majority of it is dark grey, with common anhydritic shale relics. Near top, they form a continuous network, enclosing crystals and nodules of salt (Figure 94). The shale contains, particularly along the shale-halite interfaces, finely crystalline pyrite. Soon shale becomes dominant, leading over into the Camillus formation.

Figure 93

Dark shale bed, overlying zone of white anhydrite nodules and containing discontinuous laminae of light grey argillaceous anhydrite and bright crystals of halite

Actual size

Upper Syracuse, Watkins Glen, New York

Figure 94

Dark salt containing matte intercrystalline shale network

Actual size

Upper Syracuse, Watkins Glen, New York

Fig. 93

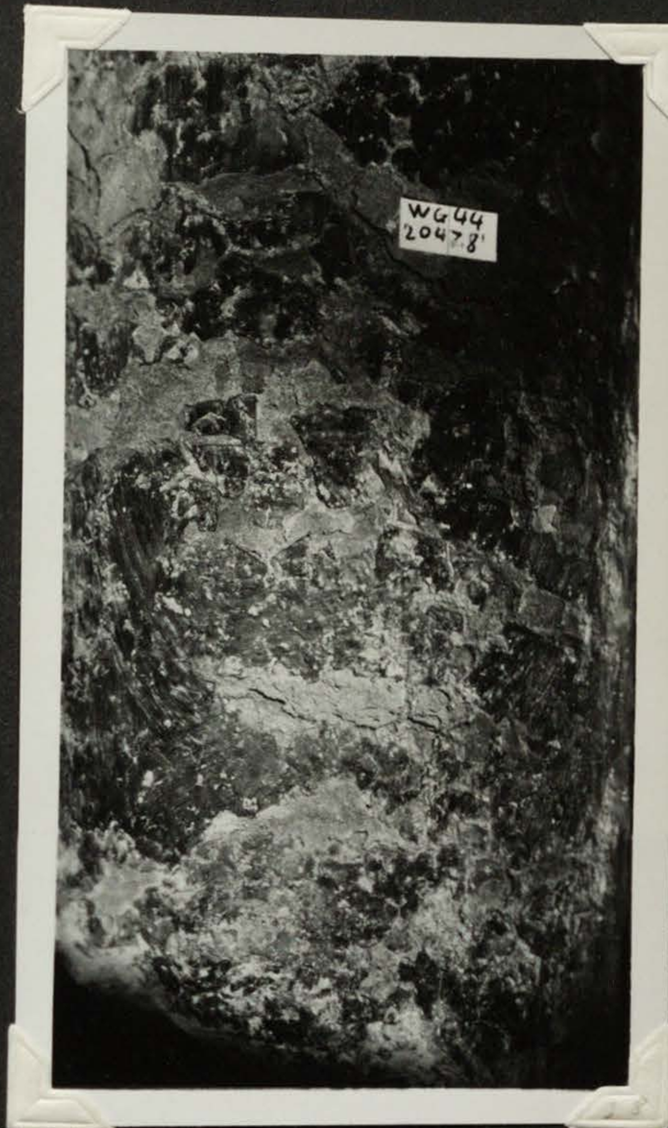


Fig. 94

CAMILLUS FORMATION

The shales of the Camillus Formation overlie the salt-bearing Syracuse Formation all over the Appalachian Basin. To the south, the Camillus lies over successively higher salt horizons, the shale interbeds between which merge into its base as the uppermost salt beds wedge out toward the north.

The Camillus Formation is a rather uniform sequence of dark gray shales, anhydritic in its lower, and increasingly dolomitic in its upper portion. With gradual increase in its carbonate and decrease in its clay content, it passes gradually into the overlying Bertie dolomites.

Only the lower, more argillaceous, anhydritic portion of the formation has been cored. This section is characterized by shale contents fluctuating from 60 to 85%, except at its base where the proportion of halite nodules is still considerable.

The base of the formation displays a thin gradational sequence. Near the top of the Syracuse Formation, salt nodules are enclosed by a gradually thickening network of anhydritic shale. Eventually, this anhydritic shale becomes dominant, and the single crystals and nodules of dark halite, embedded in the shale and surrounded by pyrite, become infrequent. Finally, halite occurs only in anhydrite laminae, forming long thin lenses; only in the most argillaceous, least anhydritic shale layers does the halite occur in the shale itself. In the latter case, under the microscope, small halite lenses (50 microns) occur in the less bituminous laminae.of the shale. These lenses often join to form discontinuous laminae; less often, vertical structures of similar shape occur. In both cases, presumably, the salt was redistributed along minor channels in the shale; in the case of the horizontal structures, the bituminous-argillaceous laminae served as obstructions in the path of the saline solutions, restricting them to the less bituminous laminae.

Particularly in its lower part, the shale is often well laminated. Under the microscope, flat laminae of quartz silt and fine sand are often observable. Dark thin laminae of somewhat anastomosing orgainc matter enclose lenses of less bituminous and possibly more dolomitic shale.

In the most argillaceous layers (shale ratio 90%) some of the most bituminous shale laminae are broken up into randomly dispersed microscopic fragments,often isometric, occurring in random orientation in the less bituminous matrix. Less bituminous shale fragments also occur.

The shale fragments are, under the microscope, accompanied by abundant small, often euhedral crystals of pyrite (0.02-0.06 mm). This pyrite is more common in the more bituminous laminae and fragments, than in the less bituminous matrix.

The most prominent feature of the shale, however, is the abundanee (3-30%) of lenses and plicated laminae of anhydrite. Most of these are pure, only a few contain shaly admixture. The shape of the smaller lenses (up to 1 mm thick and a few mm across) often resembles crystal forms (gypsum?), with horizontal long axis.(Microphoto 95). The larger lenses (0.5-3 mm, rarely 8 mm thick and 1-3 cm long) are frequently mildly undulating (Fig. 96, 97).

Stronger plications are often displayed by the almost continuous nodular beds (Fig. 98), and also by the continuous laminae, 1-8 mm thick. In the most argillaceous layers, some of the small lenses are coated by a thin lining of pyrite crystals replacing anhydrite.

The anhydrite content of the shale varies within cycles 5-11 ft thick. Often, high anhydrite content appears abruptly over shales containing little or no anhydrite, and decreases gradually to be overlain by the highly anhydritic base of the next cycle. Less commonly, the anhydrite content increases and decreases at equal rates, forming symmetrical cycles.

On the bedding planes of the shale, a polygonal network of minor mudcracks is often present; the cracks are filled with anhydrite. Connecting the anhydrite lenses, these cracks often form a vertical anhydrite network in the shale (Fig. 99).

The anhydrite filling the fissures differs considerably from that of the lenses and laminae. It consists of larger, euhedral crystals, normally perpendicular to the walls of the fissures. Besides anhydrite, secondary dolomite and halite occur as fracture filling. The dolomite forms large rhombohedra (1 mm), often incorporating or coat-

Figure 95

Photomicrograph of grey shale, containing lenses of anhydrite (white) and flakes of carbonaceous matter (black; possible algal chips)

Nicols crossed; Magnification 12 diameters

Camillus Formation, Watkins Glen, New York

Figure 96

Light grey lenses of argillaceous anhydrite and short steep cracks filled with white anhydrite in grey shale

Actual size

Camillus Formation, Watkins Glen, New York

Figure 97

Anhydrite lenses, grading into intraformational anhydrite breccia, in grey shale. Dark flakes, possibly algal chips, also occur. Note similarity between microscopic (Fig. 95) and macroscopic (Fig. 97) structures.

Actual size

Camillus Formation, Watkins Glen, New York

2024.8

Fig. 95

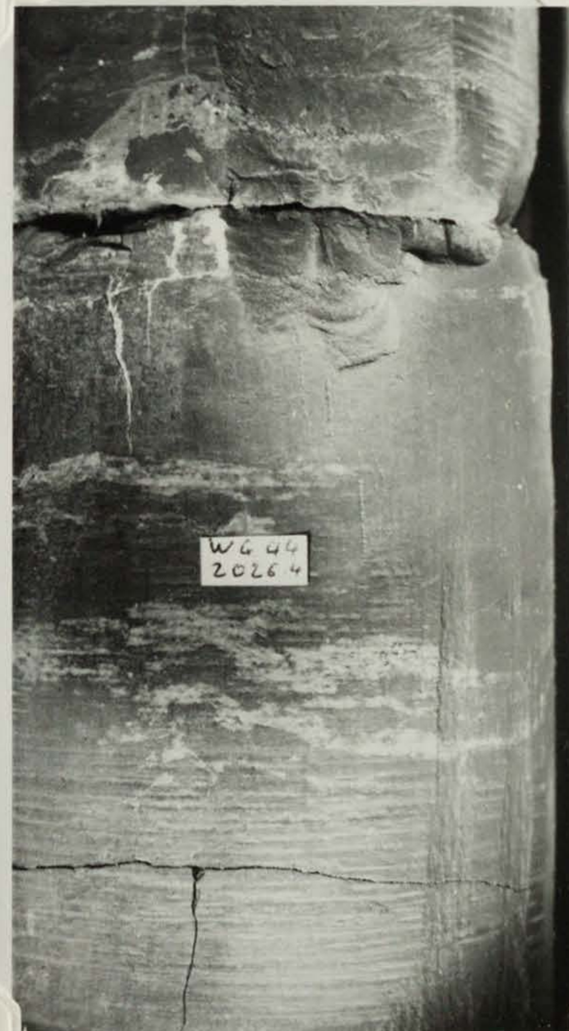


Fig. 96

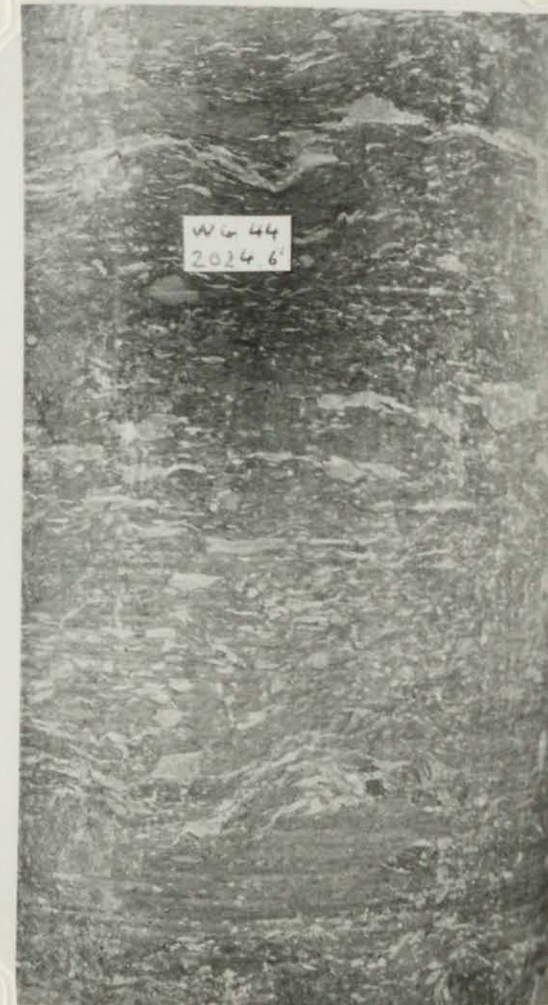


Fig. 97

Figure 98

Zone of white anhydrite lenses in grey shale

Actual size

Camillus Formation, Watkins Glen, New York

Figure 99

Anhydrite lenses, connected with short vertical cracks filled with anhydrite, in shale

Actual size

Camillus Formation, Watkins Glen, New York

Fig. 98

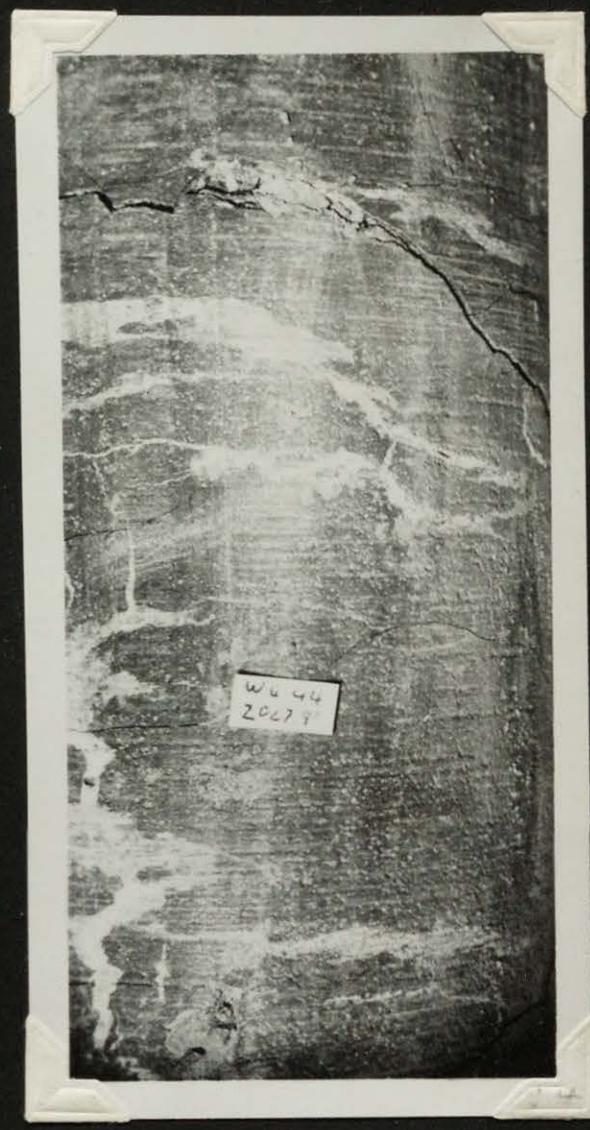


Fig. 99

ing the small dolomitic shale fragments, while the halite fills the central part of the fractures. Possibly, the saline solutions in the fractures first facilitated recrystallization of dolomite and anhydrite along their walls, and eventually deposited halite thereby healing the fracture. Since the fracture systems are restricted to single horizontal laminae only 1-2 cm thick, they seem to have formed shortly after sedimentation.

Facies, genesis, and cyclicity

Clastic sediments

The clastic sediments of the Appalachian Basin generally consist of green to grey, seldom black shales and mudstones, with some evenly dispersed, less commonly laminar, quartz silt. These shales are distal equivalents of the red sandstones and shales which make up the fluviatile sequences to the east, in the western piedmont of the Taconic Mts. The usually homogeneous mixture of silt and argillaceous matter is apparently due to flocculation of the clays in a saline environment; the separation of silt and fine sand into undulating laminae reflects winnowing by waves and minor currents. These coarser-grained laminae often display minor ripple-marks and microscopic cross-bedding.

Darker shale laminae, containing more organic matter, are frequently broken up into irregular fragments embedded in lighter shale. The chips often display fine lamination. Apparently, temporary desiccations were followed by renewed floods, embeddingthe desiccated flakes in the newly deposited sediment.

Except for the green varieties, the organic content of the shales is high; most of it is submicroscopic giving the slides a brown tint. The organic matter is often accompanied by pyrite.

The clay minerals are similarly submicroscopic; they consist of illite and somewhat less chlorite. In this matrix, quartz derived from both magmatic and metamorphic rocks is accompanied by flakes of muscovite.

Dolomite, anhydrite and halite are contained in the shales in approximately equal amounts. The dolomite is generally granular, finely crystalline (-10 microns), hypidiomorphic; larger quartz grains are frequently accompanied by rounded dolomite grains of similar size (30-50 microns). It appears that much of the dolomite was deposited with flocculated aggregates of clay and fine silt. The resulting pelletal structure is occasionally perceptible, with darker pellets enclosed in lighter matrix.of small proportions.

In the darker shale varieties, which contain higher levels of organic matter, the latter often forms horizontal, sometimes anastomosing laminae. Where microscopic cross-bedding occurs, the coarser oblique microlaminninae alternate with such dark, organic-rich microlaminae in each set. It is conceivable that these laminae of organic matter represent algal coating of the sediment surface. In the most silty laminae a burrow-like structure has been observed (Subunit E4); apparently, there were periods when inflowing fresh waters sufficently diluted the brine to permit the appearance of bottom fauna. From the outcrops,relatively rich fauna has been described from isolated shale beds.(Alling and Briggs, 1961). Even there, however, sthenohalinic elements are extremely rare; they consist mainly of planktonic and nectonic species washed in from their normal marine habitat.

Anhydritic shale

In the shale described above, evaporite-free beds are rare. The most common evaporite mineral is anhydrite, occurring in the form of lenses and nodules. As indicated, the shape of the very small lenses occasionally resembles gypsum crystals and might be the products of replacement of small gypsum crystals lying horizontally (Fig. 100-103). The larger lenses are frequently composite ones, with their constituting elements separated by thin shale network. The largest concretions reach 10 cm, their surface, as well as that of many smaller ones, is partially coated by pyrite. Apparently, where the organic matter contained in the sediment was more than what could be subsequently oxidized by the limited amounts of oxygen contained in the interstitial brine, this organic matter has, possibly with bacterial participation, reduced some of the sulphates. The resulting sulfide ions could, then, react with the iron contained in the interstitial brines, while the calcium could be kept in solution by derivatives of the decaying organic matter.

Frequently, the flat anhydrite lenses associate to form lenticular laminae, with little but distinct shale network intervening. The longer and thinner an anhydrite lens, the more undulating it is; thin laminae are often intensely plicated and contain interfolded microlaminae of shale. Below and above these disturbed laminae, the lamination of the shale is undisturbed. Regular alternation of shale and anhydrite laminae does not occur here.

Figures 100-101

White lenses of anhydrite and black chips of organic matter in pyritic shale. The shape of the anhydrite lens below may be that of a primary gypsum crystal replaced diagenetically by finely crystalline anhydrite.

Vertical section, magnification 40 diameters

Camillus Formation, Watkins Glen, New York

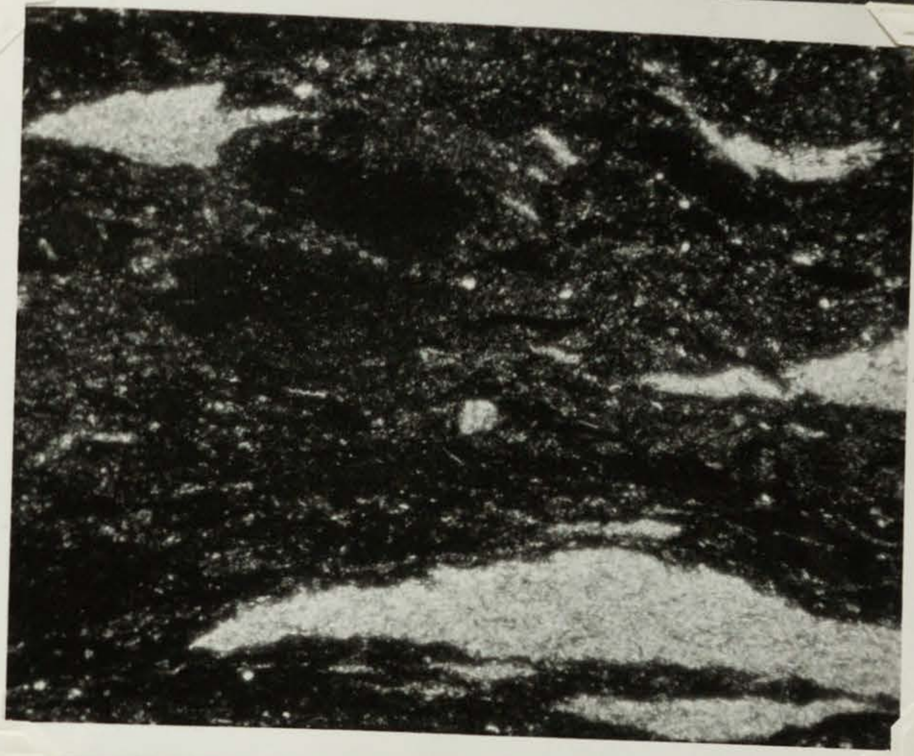


Fig.100

Fig. 101

Figure 102

Detail of large anhydrite lens from Fig. 101, showing crystalloblastic replacement structure (presumably after gypsum).
Vertical section, nicols crossed, magnification 1000 diameters
Camillus Formation, Watkins Glen, New York

Figure 103

Plicated lamina of almost opaque shale, containing finely crystalline dolomite, in anhydrite lens in shale. Near this organic lamina-remnant, the anhydrite consists of similarly plicated fibers, away from it it grades through fine matrix into coarser porphyroblasts. The plication reflects early diagenetic growth of the anhydrite nodule; replacement of anhydrite and finely crystalline dolomite, by dolomite megacrystals (grey, up to 1 cm) containing anhydrite relicts, took place later.
Vertical section, nicols crossed, magnification 40 diameters
Upper Syracuse, Watkins Glen, New York

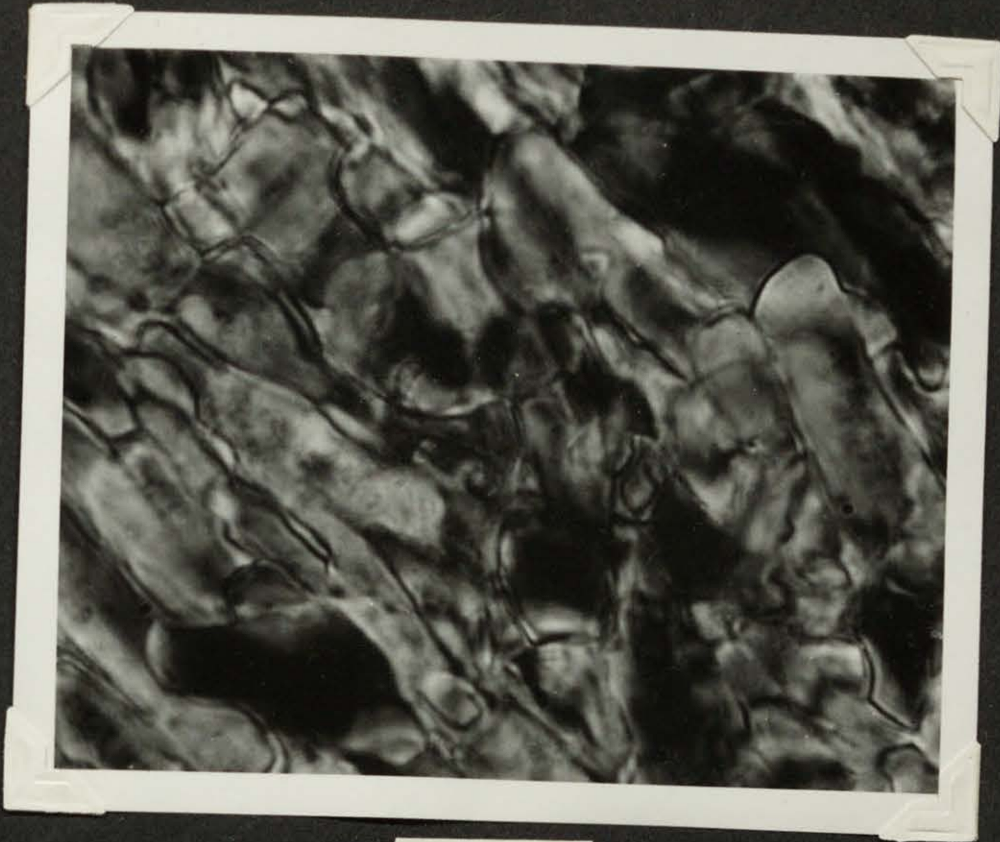


Fig. 102

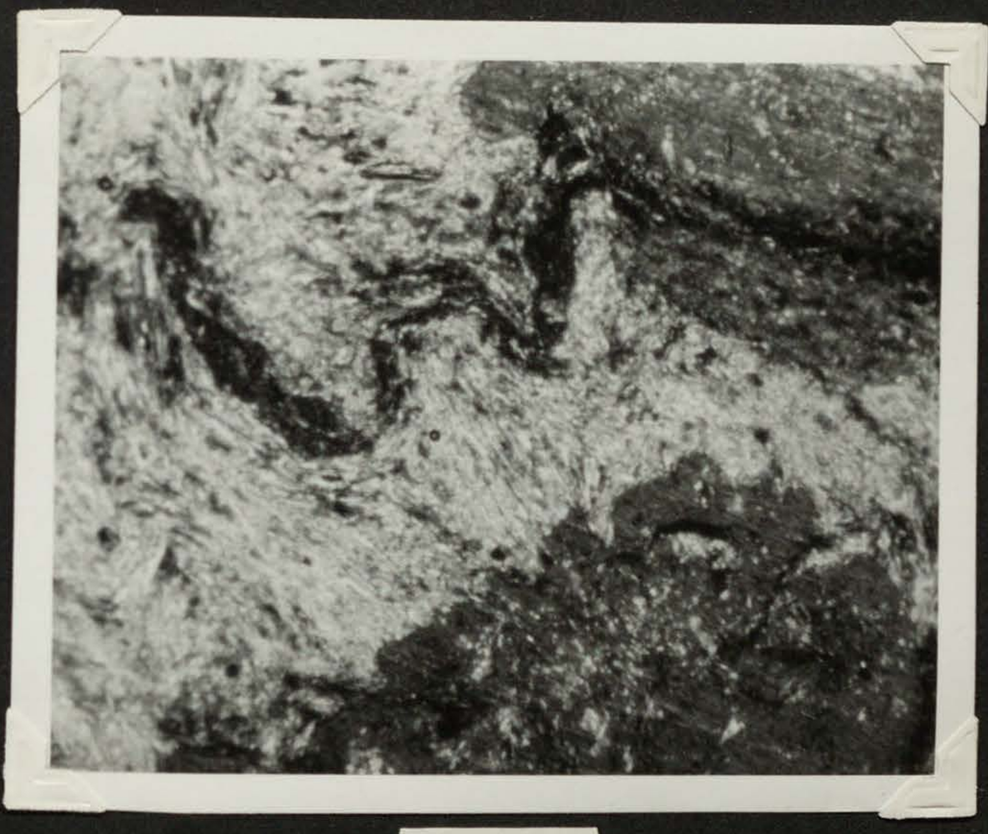


Fig. 103

Apparently, all these lenticular structures developed below the sediment surface, in an anisotropic sediment favoring horizontal orientation. Some or all of the lenses, and particularly the plicated laminae, may have consisted during some (though not necessarily the initial) stage of their development of gypsum.

The most plausible source of the Ca and SO_4 ions is the interstitial brine; increasing concentrations leading to intensive precipitation may have been induced by temporary exposure of the sediment surface to subaerial evaporation. Periodic flooding may have replenished the brines in Ca and SO_4, while permanent flooding led to continued shale deposition. Development of subsurface horizons of $CaSO_4$ lenses in recent sediments has been extensively described by Strakhov (1954, 1962, Engl. 1970)from the coastal plains of Central Asia, and by Shearman (1963) and Kinsman (1966) from the Gulf Coast. Strakhov notes that the $CaSO_4$-zone extends over several thousand square miles and grades into a shallower halite horizon in the subsoil in the direction of the Caspian Sea.

Even temporary increases in the rate of evaporation, or strong offshore winds might sufficiently tilt the surface of an often detached, extensive, shallow inner sea to temporarily expose sediments over large areas. Cycles of oscillating anhydrite content in the Camillus shales, 5-11 ft each, may represent periods of desiccation. During each period of submersion, silt and clay were deposited until the water withdrew (or sediments accumulated to sea level). Then capillary waters started evaporating, resulting in interstitial precipitation of $CaSO_4$. This dry stage, with little or no sedimentation to compensate for subsidence, as well as a change in wind direction or in the evaporation/precipitation ratio, could ultimately lead to renewed submersion. Even before this, however, temporary floodings may have occurred, partly by relatively fresh waters due to increased precipitation in the mountain framework, partly by waves particularly during storms. Some of the clay flakes occurring in the shales may have been embedded during such ephemeral floodings.

Polygonal networks, commonly observable on the bedding planes of the Syracuse shales, witness such temporary desiccations. Similar structures have been described by Stewart (1954) in the shales underlying the evaporites of the British Zechstein. In the Appalachian Basin, these suncracks are filled with carbonate and anhydrite along their margins and halite in the middle, possibly indicating gradually increasing concentration of the brine remaining after a temporary flood in the cracks of the desiccated sediment.

Halitic shale

In those parts of Central Asia mentioned above, in areas further from the sea the subsoil contains calcium sulfate concretions and lenses, while closer to the sea halite appears at shallower levels. Such halitic shales occur here as well. They are permeated with halite and contain frequent halite lenses. Lenses occurring in the green shales are frequently red, indicating mobilisation of iron at higher E_H levels.

The halite lenses usually resemble the anhydrite lenses of other horizons, ranging in shape from flat lenses to irregular nodules. Their surface is also frequently coated with pyrite and organic matter. Within this coating, however, another coating frequently appears: anhydrite often coats as well as separates the halite lenses. In the Camillus shale, halite lenses frequently appear in the central section of anhydrite laminae and lenses. These structures show multiple precipitation similar to the one described above in the mudcracks of the anhydritic shale. First $CaSO_4$ precipitates from the pore brines, then, along the same channels and filling the central part of the pores, halite becomes emplaced, pressing aside the shale and anhydrite coating. This phenomenon is particularly evident in the E4 shales, where often star-shaped short (1-3 cm) pressure-cracks surround the emplacing salt lenses and crystals (Figure 44) ; the points of these stars are filled with recrystallized dolomite, andhydrite, and quartz, while their centre consists of halite. Mudcracks and irregular cracks are less common than in the anhydritic shales, possibly indicating less frequent emersion. Thicker halite beds also appear; they are frequently haematitic in the vicinity of the green shale, indicating that the iron leached from the shale was re-precipitated in a more oxidizing environment.Similar crystallization of haematite and quartz from iron and silica leached from the associated clay was described by Stewart (1951), from halite layers of the British Zechstein.

The contacts of these halite layers with the overlying shale beds are often lined with anhydrite. Apparently, these thin halite beds, correlated only for a few miles, were deposited in shallow bodies of water during periods when the detritus supply was lacking. Evidently, such beds could deposit many times, while only a few of them would survive subsequent leaching.The brine from which they precipitated was , like the interstitial brines of the halitic shales, already often depleted in $CaSO_4$, and only subsequent flooding by more sulphate-rich waters produced a thin anhydrite coating at their top.

Argillaceous dolomite and dolomitic shale

Argillaceous-dolomitic rocks in which dolomite or clay predominates in various ratios are, particularly in the upper Syracuse (Unit F) the most common non-halitic rocks. They are dark grey, and often contain small quantities of pyrite. Fine lamination is distinct. More argillaceous laminae, containing thinner, anastomosing laminae of dark organic material, alternate with less argillaceous ones. Lenses of similarly less argillaceous dolomite are contained within the more argillaceous laminae.

The ratio of organic components is generally high, and it is conceivable that the anastomosing networks themselves represent partially fossilised algal mats. Apparently, these sediments were deposited in quiet water during periods of lower, but still substantial inflow of terrrestrial detritus. Apart from the possible, and hardly estimable, influence of algal material, the fine lamination probably resulted from periodic, perhaps seasonal, deposition. During the drier periods or seasons carbonate mud deposited, while the more humid season brought in clay from the neighbouring dry lands.

The argillaceous dolomites and dolomitic shales occasionally display brecciated structures, possibly as a result of waves destroying dolomites deposited during the more queiet periods. Angular to slightly rounded fragments of pure dolomites (Figure 29). occur either randomly dispersed or only slightly dislocated in the more argillaceous matrix. The brecciated intervals are normally underlain by normal horizontal lamination. (Figure 24). Locally.

(Subunit E3) the breccia overlies a massive dolomite with a sharp, erosional contact. The top of the dolomite is intensely weathered (Figure 34): numerous thin, short fissures, filled with the overlying argillaceous matter, cut into it and die off downwards. The weathered surface is covered by what originally was loose detritus; laminae of dolomite fragments intercalate into the overlying argillaceous bed.

As in the purer shales, lenses of anhydrite are common; the small lenses associate into larger ones and, less often, the anhydrite forms plicated laminae. The genesis of the anhydrite lenses can be interpreted as that of those contained in the shale, i.e. as a process of subsurface precipitation from capillary waters evaporating under arid conditions from temporarily subaerially exposed sediments and periodically replenished by spray and intermittent wave flooding (Shearman, 1963; Kinsman, 1969; Fuller and Porter, 1969). The lower clay content of the sediment may have facilitated the process, partly indirectly, as it reflected the lessening fresh water inflow leading to higher brine concentrations, and partly directly, as the resulting higher permeability facilitated the movement of capillary brines after the sediments had emerged above the water table.

Salt-bearing argillaceous dolomite and dolomitic shale

In the argillaceous dolomites, as in the s..les discussed above, occasionally halite occupies the central part (1-2 mm) of the anhydrite lenses. If the latter precipitated from gradually concentrating interstitial brines in sediments temporarily exposed to air, the halite marks the last stage of this concentration, filling in the central part of the channelways and cutting off further brine supply.

In other cases, rare thin laminae and flat lenses of salt intercalate with the laminated argillaceous dolomite, while the bedding planes often display a polygonal network of thin cracks filled with halite. Some of these structures may have formed on the sediment surface, particulary as halite crusts on top of temporarily exposed sediments during periods of temporary withdrawal of the lagoon. If the new flood was saturated in NaCl, these crusts remained partially or fully preserved.

Such desiccational phenomena are even more conspicuously displayed by those beds of argillaceus dolomite which contain several mm long, thin, curved flakes of finely laminated organic-rich shale and similar flakes of halite containing some finely dispersed dolomite. Though some recrystallization (secondary dolomite) occurs in the halite flakes, their boundaries are perfectly sharp. The flakes tend to lie horizontally, but their axis mildly oscillates; flakes of different material (halite and shale) occur side by side in the same argillaceous dolomite lamina.(Fig. 104-105).

Figure 104-105

Flakes of dolomitic salt in argillaceous dolomite. Together with finely laminated flakes of argillaceous dolomite, (Fig. 105), flakes and irregular fragments of dolomitic halite, often arcuate and sometimes slightly rounded, are embedded in the argillaceous dolomite matrix. Halite (black) constitutes the major, inner part of the flakes, fine-grained dolomite (white) cemented with halite is concentrated along their margins. Apparently, the replacement of dolomite by halite took place prior to the formation of the flakes, possibly in the desiccating surface layer of the sediment.
Nicols crossed, magnification 40 diameters. Vertical section.
Upper Syracuse, Watkins Glen, New York

Fig. 104

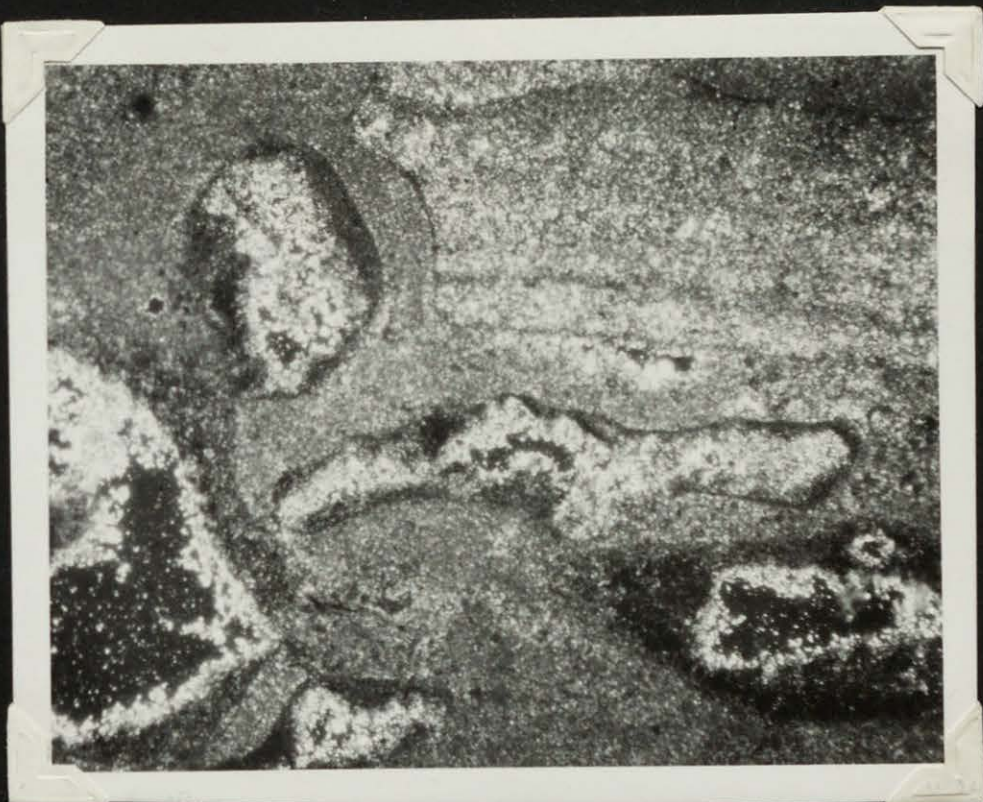


Fig. 105

Apparently, these curled flakes were formed by replacement of dolomite by halite in the surface layer of the temporarily subaerially exposed sediments, followed by continued desiccation, disintegration and embedding upon renewed flooding. As before, the brines flooding the lagoon must have been saturated in NaCl, otherwise these thin flakes would not have survived redeposition. The fact that the replacement was by halite and not by anhydrite already indicates a high $NaCl/$ $/CaSO_4$ ratio in the brines carried over the sediment surface by wind spray and wave flooding.

The motion of these shallow, agitated lagoon brines has locally led to the formation of oolites (Subunit F2/3). In sets of alternating more argillaceous and more dolomitic laminae, the latter contain in alternating zones clear halite and dark dolomite oolites, both about 0.3 mm large and often flattened at their bottom. Some halite oolites have a dark core, others display a dolomitic, irregular rim. The dolomite oolites, from which they may have formed, are stained dark with organic matter and contain a lighter core; some of them are spastolithic (the crust is broken into the core). Apparently, the ooliths were deposited in soemwhat agitated water, while the intervening laminae containing argillaceous matter mark deposition in quiet water. Alternation of laminae with halite and laminae with dolomite oolites would indicate that the replacement of aragonite oolites by halite followed their formation very soon and may even have been a seasonal phenomenon (Fig. 106-107).

Figures 106-107

Flattened, frequently broken ooliths of halite (white) and bituminous dolomite (black; bottom of Fig. 107) in anhydritic dolomite matrix. In alternating laminae, the ooliths consist predominantly of halite and predominantly of dolomite. Some of the halite ooliths have a dark core, others contain dolomite along their rim. Possibly double replacement (aragonite ooliths by dolomite and dolomite by halite) is involved.

Nicols not crossed, magnification 40 diameters

Upper Syracuse, Watkins Glen, New York

Fig. 106

Fig. 107

Figure 108

Dark dolomite ooliths with lighter core. Anhydritic dolomite matrix has been largely replaced by secondary large single dolomite crystal.

Nicols not crossed, magnification 160 diameters. Vertical section.

Upper Syracuse, Watkins Glen, New York

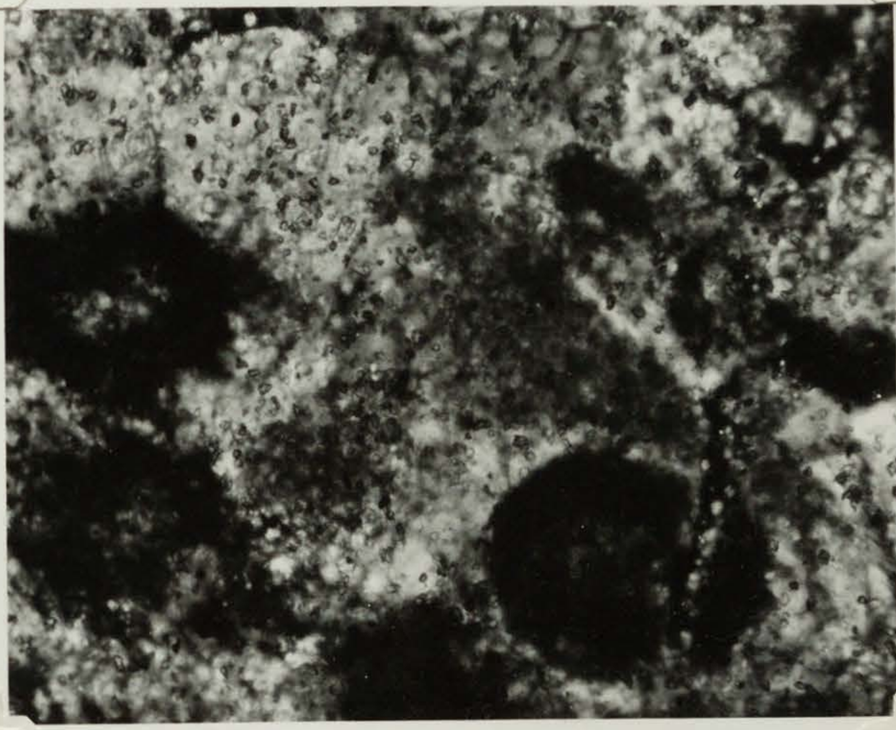


Fig. 108

In this context it is interesting to recall the horizons of armoured shale balls, up to 30 cm large, which occur in the Vernon salt of Retsof, New York; their presumably avalanche-like growth was probably due to similar wave action. Possibly similar rolling, on the surface of the sediment, was responsible for the vague concentric structure observable in the dark dolomite oolites of the rock discussed above. (Figure 108).

Apart from such halite crusts, laminae, flakes, and oolites in the majority of cases the halite forms more or less isometric lenses, cm-large single crystals, or larger, round, almost isometric nodules in the argillaceous dolomite (Subunit D1, Figure 25; Subunit F1-1/2, Figure 58; Unit F in general). These halite inclusions are frequently accompanied by anhydrite lenses and sometimes also by finely dispersed pyrite. Increasing frequency of halite nodules, lenses and crystals characterize the gradual transitions, particularly upwards, from argillaceous dolomite or dolomitic shale into halite containing only networks or relics of argillaceous dolomite. These halite lenses cut across the primary lamination of dolomite laminae and across anhydrite lenses alike. Some of them may have formed as concretions in the capillary zone of subaerially exposed sediments, in similar conditions as the anhydrite lenses, but presumably at much higher crystallization pressures. Subsequent partial leaching, and wave action during renewed flooding, may have increased the disturbance of the lamination.

Particularly at the top of halite beds, similar disturbances are related to their temporary exposure before being coated by the overlying shale bed. Partial recrystallization was, in these cases, accompanied by partial leaching of the halite and filling of the intercrystalline spaces with the still plastic clay.

Dolomites

Non-argillaceous carbonates form a significant portion of the Syracuse Formation. They are particularly common in the Middle Syracuse Dolomite Suite (Subunits E1-3), and in the lower parts of the Lower and the Upper Syracuse (Subunits D1/2 and F1/2). Most of them are gray or brown, finely crystalline dolomites, consisting of hypidiomorphic grains 2-10 microns in size. Among the dolomites, stromatolitic, inorganically laminated, and massive structures can be distinguished.

1. Algal stromatolites

Diverse stromatolitic structures form a substantial part of the Syracuse Formation. They occur in extensive beds, up to 5 ft thick. One of these beds in the F2 subunit can be traced in the Cleveland Mine for several miles without significant changes. Due to these excellent opportunities for observation, this bed will be described in greater detail.

The algal rocks of this bed develop from a shale bed containing anhydrite concretions and overlying a thick salt layer with a sharp contact. The salt below the shale bed, and in vertical discontinuous cracks within it, has a reddish hue (Fig. 109). In the shale, anhydrite concretions, shale flakes, and algal chips are common.

In the overlying grayish brown dolomite, three major horizons can be distinguished. The basal horizon (0.6 ft) directly overlies the shale; it is still argillaceous and grayer than the upper dolomite zones.

Figure 109

Basal shale bed of dolomite suite overlying salt. The top of the salt is red-coloured and displays disturbed lamination. The shale contains white anhydrite nodules, dark, possibly algal chips, and discontinuous steep fissures filled with pink salt.

Height of picture: 4 ft

Upper Syracuse, Cleveland, Ohio

Figure 110

(Upward continuation of Figure 109)

Biostrome developing from basal shale bed of dolomite suite. First buff dolomite laminae intercalate with the shale, then unlaminated dolomite becomes exclusive. Near top, alternating dolomite laminae differing in their organic content display flat domal structure.

Height of picture: 4 ft

Upper Syracuse, Cleveland, Ohio

Fig. 109

Fig. 110

Its structure is marked by indistinct argillaceous laminae and anhydrite lenses, decreasing upwards with the decreasing clay content.

The characteristic algal lamination appears in the second zone (about 1.4 ft), where thicker light and thinner bituminous dark dolomite laminae alternate (Fig. 110). These laminae form upwards convex flat domes, about 2-3 ft in diameter, in which each successive lamina is slightly more convex than the one below. The laminae continue across the boundaries between neighbouring domes. Sometimes the bituminous laminae have a saw-shaped structure superimposed upon the gently convex general pattern. Anhydrite lenses and irregular cavities filled with anhydrite are common.

In the upper zone the flat domes grade into highly irregular, steeply domal, hemispherical structures.(Fig. 111,112). In a 2 ft wide hemispheroid, the height of the archcan exceed 0.7 ft for each lamina. The central part of the hemispheroids is often unlaminated and contains many cavities filled with halite more often than with anhydrite; some of these cavities are more than 15 cm across. Radial fissures, filled with halite and anhydrite are common, they proceed from the basal zone of the reef rocks to the top of the hemispheroids, emphasizing their structure. The upper "crust" of the hemispheroids is finely laminated; along their steep sides the alternating brown and bituminous dark laminae are almost vertical. The troughs between adjacent hemispheroids are V-shaped and very narrow; the laminae here cannot be traced from one dome into the other. Hemispheroids with chaotic core structure occur beside as well as on top of each other.

Figures 111-112

Upper, hemispheroidal zone of buff dolomite biostrome. The chiefly algal hemispheroids form layers alternating with layers of detritic dolomite (fine dolomite silt to dolomite breccia). Each hemispheroid consists of a structureless core, surrounded with a finely laminated crust; in the latter, laminae of higher and lower organic content alternate. The steepness of the laminated crust along the sides of the hemispheroids indicates strong wave action during growth. Radial cavities along the axis and in the core of the hemispheroids are filled with white anhydrite.

Height of pictures: 6.5 ft (Fig. 111); 3.5 ft (Fig. 112)

Upper Syracuse, Cleveland, Ohio

Fig. 111

Fig. 112

A few of them, at the top of the upper zone, are entirely surrounded by the laminated crust, thus turning into horizontally flattened spheroidal balls.(Figure 113).

The irregular, tumorous surface of the hemispheroidal zone is overlain by a thick salt layer (Subunit F2-2) with anhydrite laminae. The salt overlies the bulging hemispheroids transgressively: the lowest salt beds and anhydrite laminae wedge out along the flanks of the hemispheroids, while the higher ones gradually cover them (Figure 113).

Studies in recent and ancient sediments by Black, Monte and Logan indicate that such structures were formed by blue-green algae in the intertidal zone. The flat, sinusoidally continuous laminae represent algae active in low-energy, protected intertidal environment; the successively more bulging hemispheroids and the discrete spheroids result from increasing exposure to waves in the lower part of the intertidal zone. Covering by the salt laminae in transgressive manner indicates that the intertidal hemispheroids were ultimately flooded by a supersaturated sea.

Thus the following sequence of events can be postulated:

1. Withdrawing sea exposes coastal salt layer to subaerial weathering and recrystallization.
2. Periodic terrestrial floods deposit argillaceous mud over the coastal salt plain. Incursions by the sea are rare. Evaporating interstitial brines deposit $CaSO_4$-concretions in the capillary zone.

Figure 113

Bedded halite overlying hemispheroidal zone of biostromal dolomite suite. The algal hemispheroids consist of massive core and laminated crust; in the latter, fine laminae contain alternately more and less organic matter. Locally, the hemispheroids become flattened spheroids, indicating very strong exposure to waves during growth. In the overlying halite, beds of halite (H) alternate with beds of anhydrite (A). The lowest anhydrite and halite beds wedge out along the flanks of the tallest hemispheroid; the third anhydrite bed is the first to transgress it.

Height of picture: 3 ft

Upper Syracuse, Cleveland, Ohio

Fig. 113

2. Gradual subsidence results in regular tidal floodings. Decreasing detritus inflow, possibly due to increasing submergence of the source area, favors the growth of algae.
3. Increasing submergence exposes algae to stronger waves, inducing them to build protective structures. Salinity of flooding sea water rises as the pseudomarine zone diluted by terrestrial inflow retreats.
5. Algal hemispheroids become gradually covered by permanent brines, depositing evaporites first over the lowest, then over successively higher hemispheroids, ultimately covering the highest structures.

Similar stromatolitic sequences are common also in the eastern belt of the basin (Watkins Glen), though here the size of the cores limits their observation. In the Lower Syracuse (Subunit D1/2), a thicker shale sequence overlies the salt and grades upwards into laminated argillaceous dolomite. Over this a stromatolitic layer (about 3 ft) follows. In its lower portion the alternating fine bituminous dark and bitumen-poor light laminae become increasingly undulating (Figure 12); in the upper, hemispheroidal portion, vertical elements as well as salt-filled fissures and cavities complicate the lamination (Fig. 13). Near the top, contorted laminae are partially impregnated by salt; alternation of cavernous and compact laminae is characteristic (Fig. 114-115). Here, too, the stromatolitic layer is eventually overlain by salt (Fig. 116).

Another thicker stromatolitic layer in the east occurs in the lower part of the F1/2 dolomite, i.e. below the F2 salt which

Figures 114-115

Laminated biostromal bed in dolomite. Pores and cavities are filled with dark salt.

Actual size

Lower Syracuse, Watkins Glen, New York

Figure 116

Cavernous biostromal laminae at top of dolomite biostrome grading upwards into massive dolomite. The dolomite is overlain by halite containing anhydrite film among the crystals. Steep fractures and cavities in the dolomite are filled with dark salt.

Actual size.

Lower Syracuse, Watkins Glen, New York

Fig. 114

Fig. 115

Fig. 116

contains the biostrome layer so well exposed in the Cleveland Mine. At Watkins Glen, this layer is about 3 ft thick; like the biostromes described above, it, too, displays an upwards increasingly complex lamination. (Figures 69 and 70). The very irregularly laminated hemispheroids at top are overlain by quiet argillaceous dolomite laminae.

Apart from these major occurrences, thinner biostrome beds, laminae and lenses occur in most major dolomite layer. These structures are often cavernous; the cavities are generally filled with salt (Figures 35 and 36). Most often, these algal-stromatoporoidal structures occur close to contacts between thicker argillaceous-dolomitic and overlying salt layers, i.e. in positions similar to those of the major biistromal layers.

Apparently, the massive development of **algal growth** is controlled by the clay content, salinity and depth of the water: conditions become optimal when clay inflow decreases and, with the decrease of terrestrial inflow, the salinity rises to slightly superhaline levels. Their position indicates that part of them may have formed in pseudomarine conditions (Sloss, 1953) in a zone sufficiently close to the delta to permit the small inflow of terrestrial waters to dilute the highly concentrated brines to normal marine concentrations, but sufficiently far from it to avoid brackish conditions or high clay contents. Gradual advancement of the more concentrated basin brinestowards the shore ultimately encroaches upon this zone, terminating algal activity. Similar relationship was found between algal activity and evaporite deposition in the pre-Ochoan evaporites of Texas (Stewart, 1954).

2. Clastic dolomites

In non-biostromal laminated dolomite sequences, layers, beds and laminae of fine, often gray dolomite alternate with similar structures of silt-to very fine sand-sized dolomite. Together with the coarse dolomite grains (30-80 microns), quartz grains of similar size often appear, indicating common transportation. Pyrite and thin laminae of organic matter occur (Fig. 117).

The coarse dolomite layers overlie the fine ones withsharp, erosional contacts, reflecting deposition from currents cutting into sediments deposited in quiet water. Where the fine laminae of coarse buff and fine grey dolomite alternate, occasionally burrows penetrate from a coarse lamina deep (up to 10 cm) into the underlying set, crossing several coarse and fine laminae and filled with coarse material.(Figure 37). Where microscopic laminae of coarse and fine dolomite alternate, the latter often contain anastomosing microlaminae of organic matter, while the coarse laminae contain quartz grains about as large as the dolomite grains. The boundaries between the coarse and fine laminae are sharp, organic-rich argillaceous partings, often coated with fine pyrite.

These carbonate silts were apparently deposited in shallow, agitated water, not very far from algal biostromes. Much detritus, particularly the coarse component, was derived from algal structures abraded by the waves. In quiet water the fine component dominated;

occasional winnowing by waves and currents produced irregular lamination, while lasting deposition from high energy environment, waves or winds, resulted in the formation of unlaminated coarser beds. Apart from these processes, an unknown proportion of the fine component may have been produced by primary chemical-biological precipitation.

In these laminated carbonates, halite is a common constituent. Its proportion oscillates in successive laminae, thereby emphasizing the lamination. In some laminae, halite becomes dominant and contains dispersed dolomite grains. Presumably, these predominantly halitic laminae were deposited during warmer or drier seasons, while the carbonate laminae of the colder or more humid seasons contain little or no halite.

3. Desiccational structures

In the thicker dolomite beds, desiccational phenomena are common. At the argillaceous dolomites, we have referred to a massive dolomite layer displaying strong weathering and overlain at a sharp contact by argillaceous dolomite containing pure dolomite fragments. The top of the pure dolomite bed is irregular, and it is through and through penetrated by downward tapering fissures filled with argillaceous material (Figure 34). Similar desiccational phenomena are displayed by a dolomite layer, overlying a biostromal dolomite bed above the uppermost Syracuse salt in the Cleveland Mine, and overlain by shale at a very sharp contact.

From this sharp, erosional dolomite-shale contact, numerous small cracks penetrate downward into the upper 0,4 ft of the dolomite, and die off gradually further down.(Figure 120). Many of the cracks have a featherlike structure, with succesively thinner secondary and tertiary cracks ramifying upwards from each "primary" crack.

Desiccational structures marked by anhydrite are extremely rare in the non-argillaceous dolomite. Underlying a salt bed in Subunit F1/2, short (-2 cm) vertical fissures are filled with anhydrite and form a polygonal network in the horizontal plane, indicating temporary desiccation.

In the unlaminated brown, often coarse dolomite, in which high porosity permitted uninhibited migration of the interstitial brines, while the unlaminated rock admitted of disoriented crystal growth, halite froms small, spheric concretions, containing finely dispersed brown dolomite grains. The size of these salt spherules is generally less than 1 mm in the east (Watkins Glen) , but increases to about 5 mm in the western, platform areas of the Appalachian Basin, where they also become more frequent. The halite spherules are often accompanied by anhydrite needles, a few mm long, dispersed in the dolomite. The anhydrite needles occur in random positions, ranging from horizontal to vertical, cutting across primary structures and incorporating dolomite grains. Though they occur in the same set of beds, the frequency of the anhydrite needles and halite spherules is negatively correlated. (Figures 118-119).

Figure 117

Coarse dolomite silt (0.03-0.08 mm)containing thin horizontal laminae of dark brown organic matter. Salt-filled pores (white) and pyrite crystals (black) are common.

Nicols parallel, magnification 12 diameters. Vertical section.

Base of Upper Syracuse, Watkins Glen, New York

Figure 118-119

Dolomitic halite spherules and dolomitic anhydrite needles in brown dolomite. The spherules (enlarged on Fig. 119) are irregular concretions, containing about 30% of evenly distributed fine (.002-.006 mm) dolomite, and displaying orthogonal cleavage system. The anhydrite needles consist of an hourglass-shaped core, containing similar finely dispersed fine dolomite, surrounded by a dolomite-free oriented overgrowth which sometimes cuts through halite spherules. Apparently, first increasing pore brine concentration in the (desiccating?) carbonate mud produced concretional cementation by Ca-sulphate and by halite, then lower concentration for a longer timeresulted in oriented Ca-sulphate overgrowth.

Section vertical, nicols not crossed, magnification 12 diameters

Middle Syracuse, Detroit, Michigan

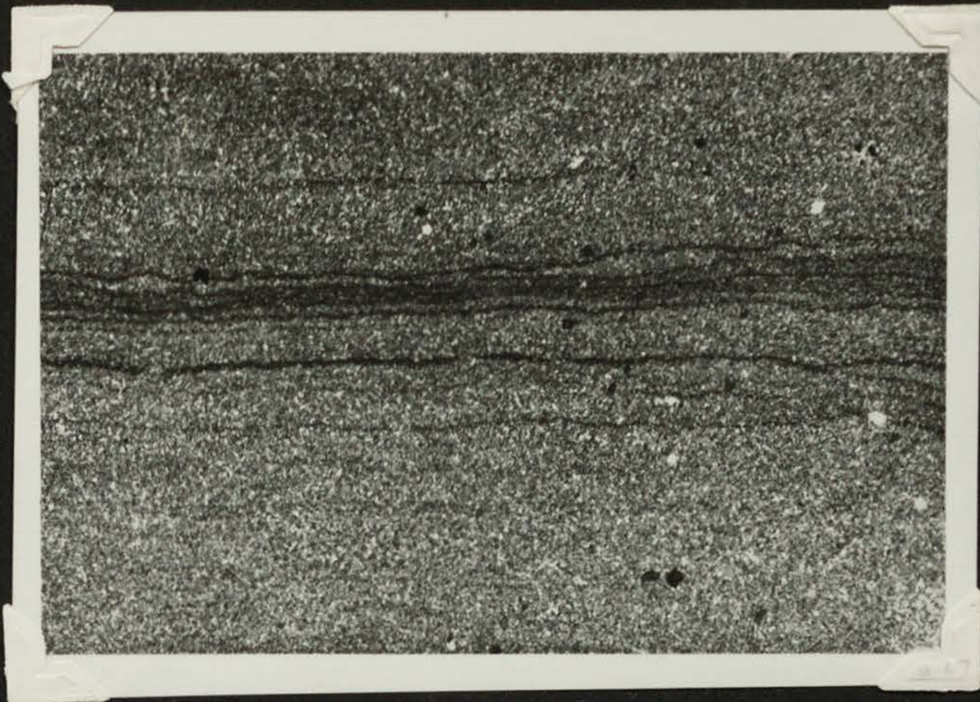


Fig. 117

Fig. 118

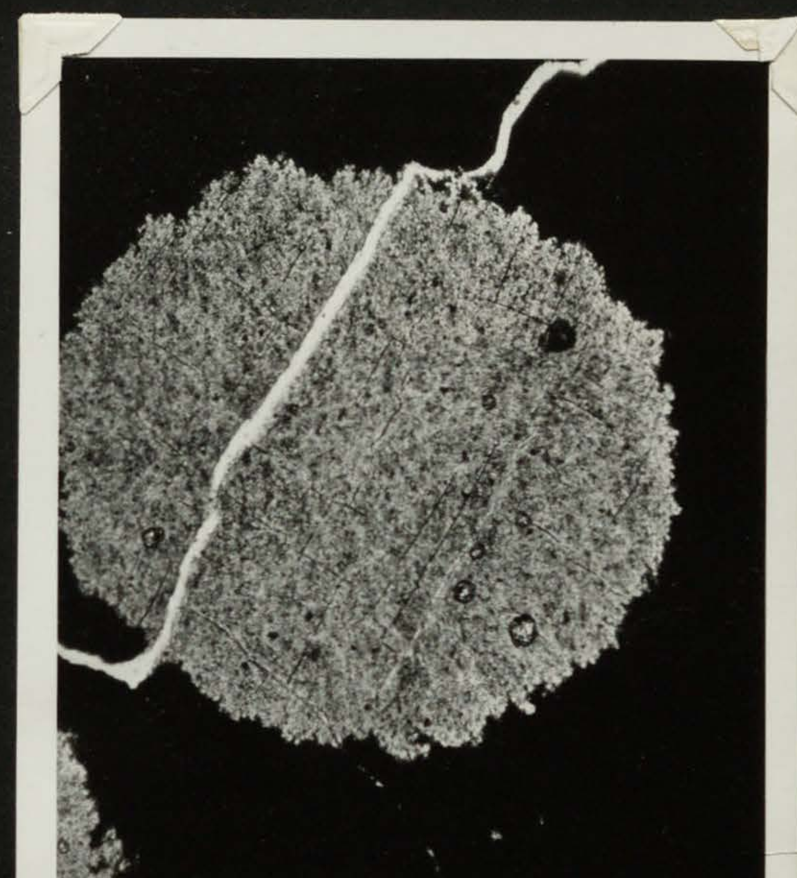


Fig. 119

Conceivably, the halite spherules and anhydrite needles were formed under subaerial conditions from evaporating capillary brines, in massive, relatively coarse, possibly wind-blown carbonate silts whose relatively isometric texture permitted disoriented precipitation. Minor nuclei, once formed, served as centres for further isometric growth. The semi-laminar zones, 1-2 cm thick, in which halite spherules are most common, may represent levels where the top surface of the capillary brine was constant for some time. Occasionally, these spherules are connected by upward ramifying microfissures, marking the route along which the capillary brine moved, (Figure 27).

A similar phenomenon is the occurrence of large halite concretions in the dolomite. Such a zone, about 1 ft thick (Figure 121), is exposed in the Cleveland Mine, about 10 ft above the top of the F2-2 salt. The salt forms irregular brown concretions, about 3 cm thick and 5 cm long, occurring in a well defined bed of the dolomite. Like the spherules described above, the concretions contain large amounts of finely dispersed dolomite. The enclosing dolomite bed is dark grey. It overlies similarly dark grey anhydritic shale at a very sharp, undulating contact; its top contact with similarly anhydritic shale is gradational. Overlying the latter shale, beds of dark shale with large anhydrite concretions follow.

Apparently, the argillaceous material was deposited by recurrent terrestrial inflows; the sharp boundary indicates possibly longer period of non-deposition. During a period when the terrestrial inflow decreased and the clayey mud deposition gave place to the de-

Figure 120

Eroded surface of partly biostromal buff dolomite (D) overlain by grey anhydritic shale (S_A). Near bottom, the dolomite consists of hemispheroidal algal domes; these are overlain by massive dolomite containing only a thin biostromal bed. Above this, the dolomite is penetrated by steep desiccational cracks, irregular; but semi-parallel and close to vertical.

Height of picture: 4.3 ft

Upper Syracuse, Cleveland, Ohio

Figure 121

Dolomite containing large amount of brown dolomitic halite concretions. The dolomite bed overlies grey anhydritic shale at a very sharp surface; upwards it grades into similar anhydritic shale without sharp boundary

Height of picture: 2.7 ft

Camillus Formation, Cleveland, Ohio

Fig. 120

Fig. 121

position of purer carbonates, the interstitial brines became more concentrated, partly because of the decreasing fresh-water inflow itself, partly because of the increasing aridity which caused it to decrease. Evaporation of these pre-condensed interstitial brines resulted in halite deposition in the capillary zone. Gradual increase in the rainfall resulted in increasing terrestrial inflow, covering the carbonates with more argillaceous deposits. In the lower beds the interstitial brines remained unsaturated, producing concretion-free sediments. In the subsequent more arid period, the interstitial brines did not reach their exceptionally high concentration during the preceding arid period, so only anhydrite concretions were formed in the upper part of the argillaceous beds (Figure 122).

Rarely, lenses, laminae and thin beds of halite are intercalated with the halite-bearing dolomite, their contacts are often marked by pyrite. Below them, the dolomite is sometimes penetrated by thin, probably desiccational, cracks filled with salt. Larger salt crystals in the dolomite become more frequent towards the halite laminae (Figure 32). These structures may indicate that minor pools depositing evaporites persisted over the carbonate mud even during periods of desiccation. Reestablishment of such pools by waves could replenish the salts in the interstitial brines as well.

Figure 122

Brown dolomitic shale overlain by grey shale along a slightly faulted contact. The dolomitic shale contains horizontal zones of anhydrite concretions.

Height of picture: 3.4 ft

Camillus Formation, Cleveland, Ohio

Anhydritic contacts between halite and shale

Although in the eastern part of the basin (Watkins Glen) pure anhydrite beds are very rare, the contact between shale and salt layers is often lined with anhydrite, a few mm to a few cm thick, particularly where the shale overlies the halite. In such cases, the anhydrite forms: large lenses separated by a thin shale network, a single continuous lamina, or a thin set of more and less argillaceous laminae, lining a somewhat undulating salt surface.

Locally (Subunit D1/2), small pockets of fragmented anhydrite in anhydrite matrix (Fig. 8) intrude into the surface of the salt; upwards they grade into argillaceous anhydrite laminae and finally into the overlying anhydritic shale layer. Apparently, the deposition of anhydrite laminae ended, and perhaps alternated with the end of, salt deposition; increasing dilution, leaching the salt below them, resulted in their collapse. Parallel increase of dilution and clay content indicates that the microcurrents leaching the salt and carrying anhydrite fragments into the pockets were concomitant with terrestrial inflow, depositing increasing amounts of fine detritus, soon becoming the dominant rock component. The detritus deposition rate often exceeded subsidence, leading to frequent desiccational processes in the sediment.

In transitions pointing into the opposite direction, i.e. from shale into halite, anhydritic contacts occur rarely. Sometimes, anhydrite lenses become more frequent in the shale toward the salt, while the latter contains anhydrite lenses at its bottom. In other cases (Subunit F1/2, Fig. 81-82), there is a continuous anhydrite lamina at the top of the shale underlying the halite. This anhydrite contains irregular fragments of shale or argillaceous dolomite, as well as crystals of salt connected by thin, salt-filled fissures. The presence of the shale flakes indicates that the formation of the anhydrite was preceded by desiccation of the shale, and randomly oriented drying flakes of the shale became incorporated in the surface anhydrite crust. The crust itself, then, formed at the shore of the now transgressing lagoon, precipitating from the same surface spray and intermittent wave flooding which had provided the material for the subsurface anhydrite and halite lenses. Eventually, the lagoon would flood these areas, compensating for the decreased detritus inflow and consequently increased surface subsidence, and deposit relatively pure halite. The rarity of anhydrite crust in such a position would reflect the observed lesser continuity of cycles in their transgressive phases (Duff et al., 1967).

Anhydritic boundary zones between halitic and argillaceous dolomite

In the preceding chapter, we reviewed the anhydritic boundary zones between halite beds and the (usually overlying) argillaceous dolomites and shales. Similar boundary zones occur along the contacts of halitic dolomite with the overlying shales or argillaceous dolomites. These zones are a few cm thick, and finely laminated. Laminae of shale, rarely also of dolomite, alternate with laminae and long thin lenses of anhydrite. Short desiccational cracks are common, they are usually restricted to single laminae (Figure 40) and form polygonal cracks on the bedding planes. An interesting feature is the local presence of randomly dispersed post-depositional large gypsum crystals (as anhydrite pseudomorphs), cutting through lamination (Subunit E1, Figure 41).

Such anhydritic-argillaceous zones, overlying anhydrite- and clay-free halitic dolomite, represent rapid change in the environment. Apparently waters of partially continental origin, carrying suspended clay and not yet deprived of their sulphate content through evaporation, flooded the often subaerially exposed dolomite silt bars and biostromes in which halite was precipitating from the interstitial brines. Probably annual oscillation in the amount of suspended clay and in the concentration of sulphates in the water resulted in alternate deposition of argillaceous and anhydritic laminae, the latter apparently the product of the drier seasons. The brine flooding the bars was shallow and desiccation was frequent, particularly in the dry season. Such desiccation resulted in polygonal mudcracks and also in the precipitation of secondary crystals from the interstitial brines. In the underlying dolomites this precipitate was halite, but here, due to the increased sulfate/chloride ratio in the brine and also to its lower concentration resulting from fresh water inflow, the secondary precipitate is primarily anhydrite. Small halite lenses occur only in a few dolomite laminae, indicating that the interstitial brine temporarily reached higher concentrations permitting halite deposition.

Halite rocks

Halite with shale and argillaceous dolomite relics

Practically all major halite layers of the Syracuse Formation contain shale or argillaceous dolomite in the form of interbeds, laminae and relics. In the more detritic lower Syracuse these consist chiefly of shale, in the Middle and Upper Syracuse mainly of dolomitic shale or argillaceous dolomite. In the upper Syracuse, they contain clay, anhydrite and dolomite in approximately equal amounts. As in the continuous shale and argillaceous dolomite beds, the anhydrite forms small (1-3 mm) lenses often resembling horizontally lying gypsum crystals. Besides dolomite and anhydrite, anastomosing dark organic-rich laminae are also common in these structures.

The halite-shale ratio oscillates within wide limits. From halite-free shale (or argillaceous dolomite) to almost pure halite several transitional phases exist, like shales with halite crystals, nodular halite with more or less congruous shale network, and halite containing discontinuous shale relics of different size. The networks and relics are randomly oriented; they are terminated by concave surfaces bearing imprints of surrounding euhedral salt crystals. Their shape does not change even if the salt crystals have, since their formation, changed their outlines and perhaps even incorporated some of the shale relics. Of the various structures discussed in the descriptive section, particular attention should be paid to Figures 88 (Subunit F4), 19 (Subunit D2/3) and 64 (Subunit F1-2).

Occasionally, the relics are coated by large secondary dolomite and a few anhydrite crystals, both euhedral toward the salt. The discontinuous relics and more continuous networks in the halite often grade vertically into continuous laminae and thin beds of shale and argillaceous dolomite. In these finely laminated beds, where purer dolomite laminae alternate with laminae containing higher amounts of argillaceous and organic matter, lenses and thin laminae of anhydrite are common. Halite lenses and crystals dissect both the lamination and the anhydrite lenses.

Some of the shale laminae overlie the halite (Subunit D2) with a serrated contact, while their top is penetrated by irregular pockets filled with salt. In another case (Subunit F4, Fig. 91, 123), an upward tapering inverted pocket, filled with clear halite, penetrates the lamina from the underlying dark salt; this protuberation cuts through the dark halite lenses contained in the argillaceous dolomite.

The salt itself is red where the shale relics and laminae in it are greenish gray (Unit D) and dark gray where the (dolomitic) shale in it is dark gray. Almost clear crystals are commonly dispersed in both varieties; locally they enlarge into small aggregates and bands, the latter almost horizontal, indicating local recrystallization probably often caused by intercrystalline brine. Similar recrystallization is often observable along the contacts of the salt with argillaceous dolomite beds; in these cases, the water for the recrystallization of the salt was partially provided by the argillaceous dolomite. In the darkest gray salt varieties, displaying the smallest amount of recrystallized crystals, gaseous

Figure 123

Clear recrystallized salt tongue, intruding from underlying salt layer into dark argillaceous dolomite containing dark salt lenses

Natural size

Upper Syracuse, Watkins Glen, New York

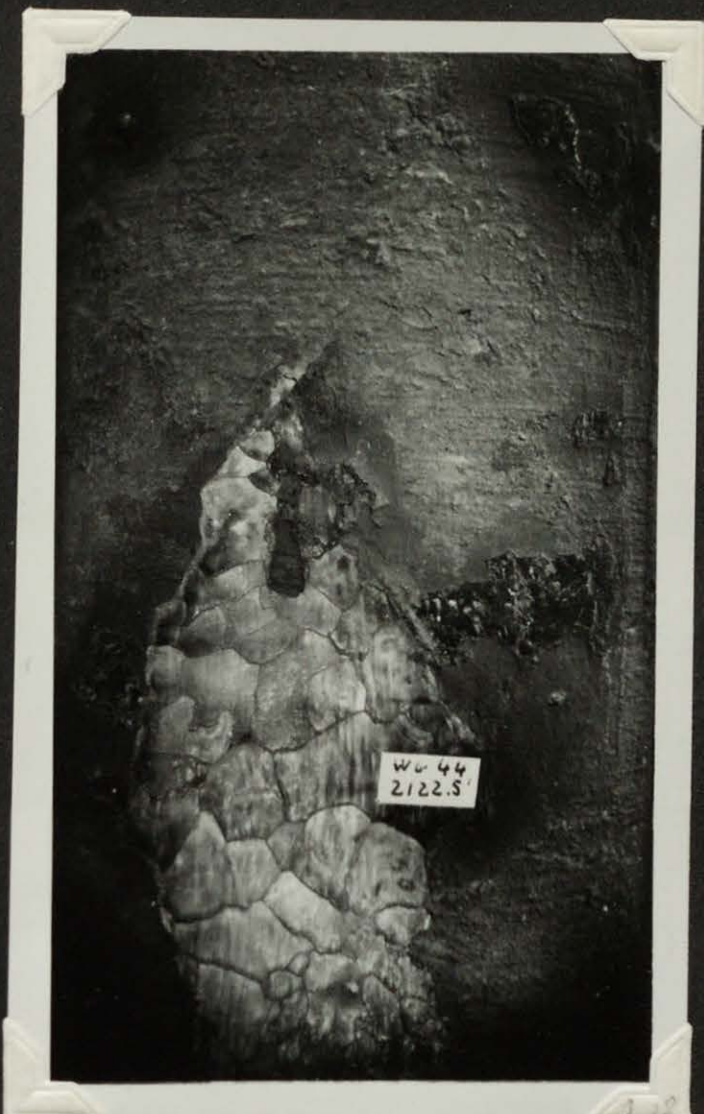


Fig. 123

H_2S is often present, emanating a strong odor upon breaking the salt.

Laminar varieties of argillaceous halite are rare. Even some of these show post-depositional replacement by halite. Thus a laminated bed in Subunit F2 consists of alternating salt and argillaceous-bituminous dolomite laminae, 3-10 mm thick. Under the microscope, the argillaceous-bituminous dolomite laminae display some quartz silt and anastomosing laminae of higher organic content. The halite laminae, on the other hand, exhibit an intricate network of slightly argillaceous dolomite. Apparently, the dolomite was here replaced by halitic solutions, which were restricted by the impermeable argillaceous-bituminous laminae to the laminae of purer dolomite. Replacement was uneven even within a single lamina.

These phenomena may indicate that even the networks and relics of shale now contained in the salt originally formed continuous laminae. Their disruption may be attributed partly to early diagenetic processes (desiccation and leaching), partly to flowage.

Temporary withdrawals of the sea, resulting in desiccation of the sediments, were apparently common, causing interstitial growth of $CaSO_4$ and halite crystals, lenses, and nodules. With increasingly frequent flooding by supersaline brines, the halite nodules grew larger and closer to the surface. These nodules gradually disrupted the shale, until only internodular shale remnants (relics and networks) were preserved.

Fig. 124

Fig. 125

Figure 124

Non-laminar top of well laminated salt layer overlain by shale. The salt contains irregular shale network and aggregates of clear recrystallized salt.

Height of picture: 4 ft

Upper Syracuse, Cleveland, Ohio

Figure 125

Laminated anhydritic salt underlying non-laminar salt of Fig. 124. Thicker halite beds alternate with thinner beds and laminae of anhydrite (A).

Height of picture: 4 ft

Upper Syracuse, Cleveland, Ohio

Recrystallization during early diagenetic weathering and leaching seems to have affected the top portion of some of the salt layers in the Cleveland Mine (Fig. 124-125). Here the major part of the salt layers is well laminated, but close to the contact with the overlying shale bed, the lamination becomes disrupted by recrystallized halite, shale networks and shale relics. Minor fissures cut into the salt from above, filled with shale infiltrated from the overlying layer.

These phenomena may be due to exposure of the salt to minor leaching, either subaerially, before being flooded by mud-laden waters, or subaquatically, as terrestrial waters dilute the superhaline waters which form a thin cover over the salt. Significantly, in the Cleveland Mine these phenomena are accompanied by the appearance of an upwards increasing reddish hue in the otherwise very light gray salt; a sign of higher oxygen ocntent in the brine in which they recrystallized. Such processes may be further complicated by the formation of salt crusts and interstitial salt crystals in the mud laminae during periods of desiccation which may have preceded, or alternated with, periods of leaching.

Recrystallization under overburden pressure is hard to separate from recrystallization during weathering: undersaturated interstitial brines in the shale laminae and beds could have similar effects in leaching, disrupting and recrystallizing the salt. The fact that in Cleveland the F2-2 salt bed shows no disruption of lamination at its bottom but shows significant disruption at its top favours the assumption of pre-burial recrystallization. Wedge-like pockets, cutting into top of the salt, also tend to support this assumption.

Recrystallization under flow, while it cannot be discounted, may not be a prominent factor in disrupting lamination: it appears that where lamination was even before flowing, even extensive flowage as in the salt domes of Germany or the Gulf Coast fails to disrupt its parallelism.

In the non-laminar halites, there is often no sharp limit between the halite lenses in the argillaceous lamina, the halite nodules surrounded by argillaceous network, and the halite of the relatively clay- and dolomite-free beds. As in the halitic argillaceous rocks, the colour of the halite is determined by the colour of the shale contained in it; in the green varieties, the iron leached from the chlorite undergoes further oxidation in the brine.

Generally, however, the redox potentials are very low; the presence of pyrite, and particularly of H_2S in the dark grey salt indicate strongly reducive conditions. This apparently is due to the organic matter deposited with the clay; in the clay-free halites, which are less rich in organic matter, reducing conditions are much less distinct. The final E_H resulted from the interaction between the organic matter and the oxygen contained in the interstitial brines; and it appears that the oxygen entering with the terrestrial waters did not survive longer transportation into the basin, while the slight wave action, withnessed by the even lamination, resulted only in little oxygen intake.

The rare fine interlamination of argillaceous rocks and halite is apparently the result of periodic, possibly seasonal deposition in the lagoon; early diagenetic replacements could take place under brine cover or in the capillary zone during periods of desiccation.

The general environment of the argillaceous halites may thus be visualized as a shallow, intermittently withdrawing extensive "lagoon" or rather margin of an inner sea to the south. The amount of decaying organic matter was very high. During periods when the climate in the surrounding mountains was so dry that no fresh water entered, the lagoon deposited salt from the inflowing sea water. Fine terrestrial detritus, hydrocarbonates and oxygen entered the basin during more humid periods, particularly from the Taconic Mountains in the east, and were partially deposited as carbonate-bearing argillaceous sediments over the halite beds. Occasional desiccations, if leaving porous argillaceous laminae in the capillary zone, resulted in the formation of first small Ca-sulphate, later progressively larger halite crystals and nodules, until the latter totally destroyed the laminar structure of the argillaceous rocks, and pushed them aside in the form of semi-continuous networks and discontinuous networks and discontinuous relics. Increasing frequency of the salt nodules reflects increasingly frequent flooding of the desiccated sediments by supersaturated brines, leading ultimately to the return of the lagoon and to renewed sedimentation in lagoonal environment.

Anhydritic halite

In the eastern part of the basin, close to the Taconic Mts., laminae, networks and relics of shale, dolomitic shale and argillaceous dolomite are the major non-halitic constituents of the halite beds, with only small anhydrite lenses contained in the shale. There are, however, throughout the Syracuse Formation, a few halite beds from which argillaceous contamination is almost absent. These layers are very light gray to almost colorless in those parts of the sequence characterized by green shale beds, as opposed to the deep dark gray or red color of most other salt beds in the eastern zone. The facies transition from shaly into anhydritic salt is gradational horizontally as well as vertically. Horizontally, towards the platform zone in the west, the clay content of the salt decreases with increasing distance from the mountainous source areas, and the shale interbeds, laminae, networks and relics increasingly give place to laminae, networks and relics of anhydrite.

These anhydrite structures are much thinner than the corresponding shale structures, indicating that decrease in the clay and carbonate rather than increase in the $CaSO_4$-supply was responsible for their formation.

The anhydrite network is only a thin, intercrystalline anhydrite film in the halite. The discontinuous relics, recorded from a bed in the Lower Syracuse (Unit D), are only a few mm thick. Here, in one of the salt beds, a system of repeatedly bifurcating, more

Fig. 126

Figure 126

Repeatedly bifurcating algal (?) tubes in organic-rich (dark) relict and underlying organic-rich lamina within light anhydritic halite. Tubes parallel with the slide show bifurcation to the right;those cutting it at an angle appear as ellipses or irregular spots. Several tubes are traceable with decreasing distinctness from the organic-rich into the organic-free portion of the salt.

Section vertical, nicols not crossed, magnification 12 diameters

Lower Syracuse, Watkins Glen, New York

or less parallel tubes consists of organic matter with dispersed grains of fine dolomite crystals; apparently they represent algal stems buried in the salt. (Fig.126;Dolomitic halite:Fig. 129).

In the east, anhydrite beds are rare in the salt. A bed of anhydrite occurs in Subunit F1-1, it is strongly dolomitic and grades upwards into dolomite. In the salt of F1-2, the anhydrite forms more or less continuous pure lenses and laminae, containing slightly argillaceous-dolomitic intervals; it overlies the salt with a sharp contact and contains flakes of shale and argillaceous dolomite. These shale, argillaceous dolomite, argillaceous-dolomitic anhydrite and pure anhydrite laminae appear to have been deposited from waters of periodically changing composition. Rare occurrences of argillaceous flakes and purer anhydrite lenses in some of the anhydrite laminae may indicate rare emergences of the sediment surface above the water table. The desiccating shale flakes could become incorporated in anhydrite which was deposited partly as crust during the desiccational period from brines brought on the surface by waves, tides and wind spray, partly from the lagoon itself upon renewed submergence.

Laminar anhydrite, so common in the platform areas of the Appalachian Basin and described from the Michigan Basin, occurs in the east only in the purest salt layers of the Upper Syracuse (Unit F). In sets 1-2 cm thick, laminae of anhydrite, a few mm thick, alternate with halite laminae of similar thickness; the latter often wedge out. Similar, though more regular oscillations in the British Zechstein, involving halite, primary andhydrite and primary gypsum, apparently reflect seasonal changes in temperature (Stewart, 1953).

Though the anhydrite laminae contain locally shale flakes and re-crystallized porphyroblasts, indicative of diagenetic growth, most of them consist of elongated laths (0.15x0.05 mm., Fig. 127) shingles with their long axes tending to parallel the bedding plane. Similar structures at Fordon (Britain) were interpreted by Stewart (1963) as indications of primary anhydrite deposition. The salt beds between the laminated sets are free of shale or dolomite relics, but often contain intercrystalline anhydrite film.

In the western, platform areas of the Appalachian Basin, farther from the detritus source areas, anhydrite laminae are common in the salt. Their thickness ranges from a fraction of a mm to about 10 cm; the thicker ones contain very thin, slightly anastomosing argillaceous-bituminous laminae. Though the anhydrite laminae locally do surround long flat lenses and wedging-out beds of salt, they are very persistent: the same laminae can be followed for several miles in the Cleveland Mine while their thickness, as well as the thickness of the intervening salt beds, stays constant with only local oscillations.

Some of the anhydrite laminae have indistinct contacts, particularly along their base. Few mm long walls extend from them in between the crystals of the underlying (sometimes also the overlying) salt bed. Such features may indicate that salt and anhydrite were deposited alternately from the lagoon, and the soft anhydrite (or gypsum) mush penetrated into the spaces between the bordering halite crystals. Subsequent recrystallization of the salt, pushing the Ca-sulphate aside, may have contributed to the phenomenon. (Fig. 128).

Figure 127

Anhydrite laminae in clear halite. In the (discontinuous) anhydrite laminae, large amounts of anhydrite porphyroblasts lie flatly or randomly oriented in a matrix and among lenses of finely crystalline anhydrite.

Section vertical, nicols not crossed, magnification 12 diameters.

Upper Syracuse, Watkins Glen, New York

Figure 128

Anhydrite lamina, disturbed by minor salt recrystallization, in clear halite. At the top of the lamina (the apparent flat top is the end of the slide), at its bottom, as well as in pockets deepening from the anhydrite to the halite, and between the halite crystals, the anhydrite is clear and consists of randomly oriented needles. The central zone of the anhydrite lamina, however, is dark; in it, organics and fine-grained dolomite are penetrated by irregular anhydrite crystals.

Nicols not crossed, magnification 12 diameters, section vertical

Upper Syracuse, Cleveland, Ohio

Fig. 127

Fig. 128

The depth of the lagoon, in which the anhydrite-bearing salt was deposited, is indicated by the fact that the lowest beds of the F2-2 salt in Cleveland, with the intercalating anhydrite laminae, wedge out on the flanks of the higher hemispheroids (see Dolomites). This would indicate that, at least in the initial stages of the evaporite deposition, the lagoon was less than 1 m deep, although it extended to hundreds of miles.

The major factor controlling the formation of interlaminated anhydrite and halite was the periodic inflow of undersaturated water (terrestrial and/or marine) into the lagoon, depositing Ca-sulfate and occasionally clay. (As in the case of the anhydrite beds described above, Ca-sulfate may have continued tc precipitate in and aroung drying clay laminae during short periods of desiccation, while some of the shale flakes became embedded in anhydrite during renewed flooding.) More saturated brines, possibly corresponding to hotter, drier seasons, produced, on the other hand, halite laminae which partially survived the next undersaturated inflow. A major feature was the very low proportion of clay and HCO_3 in the terrrestrial waters; if they had been present in larger amounts, as they were through much of the Syracuse Time, then instead of anhydrite laminae containing only a few shale flakes, shale laminae with anhydrite lenses would have been formed. The anhydritic salt is thus in all its facies (relic, laminar and bedded occurrence of anhydrite) the facies equivalent of some shaly-dolomitic salt, and marks areas and periods characterized by very low levels of terrestrial detritus inflow. Its temporary appearance in the east marks periods of maximum aridity in, or inundation of, the detritus source area, while its appearance horizontally signifies increasing distance from it. Thus anhydritic halite becomes increasingly common, argillaceous halite increasingly rare, towards the western, platform regions of the Appalachian Basin.

Dolomitic halite

In the dolomitic halite, laminar and nodular varieties can be distinguished. The laminar ones are comparatively rare. A transitional case, mentioned also at the anhydritic dolomites, occurs in the lower part of the Lower Syracuse. Here, apart from relics and laminae of anhydrite, wavy laminæ and irregular relics of organic matter, containing fine dispersed dolomite grains, occur in the salt; systems of often straight, repeatedly bifurcating tubes, often with hollow, halite-filled interior, lie semi-parallel to the lamination. In a bed in the lower part of the Upper Syracuse (Subunit F2), anhydrite is absent, and the fine lamination of the salt is provided by dispersed fine dolomite grains. Erosional interfaces, with steep channels and extremely rugged microrelief, can be commonly observed under the microscope; each erosional surface is lined from below by a thin band in which dolomite grains are more common in the salt. Algal structures, consisting of straight and slightly curved stems, with frequent bifurcations, are common (Fig. 129), characterized by high frequencies of dolomite grains. Turbulences and dragging produce straight or curved bands with chevron-like internal structure. These bands sometimes follow the lamination, sometimes line erosional micro-channels (Fig. 130).

Apparently, these laminar dolomitic salts represent lagoonal environments of salt deposition. The erosional surfaces indicate periods of leaching, which led to relative increase in the concentration of dolomite below these surfaces. The algae were probably carried by currents into the lagoon; the chevroned bands were

Figure 129

Laminae of halite, alternately rich (dark) and poor (light) in fine dolomite grains.Early diagenetic leaching produced step-like erosion surfaces and micro-channels (below centre). Long straight dark line, cutting across much of the picture at a small angle to the lamination, may be an algal stem.

Section vertical, nicols not crossed, magnification 12 diameters

Upper Syracuse, Watkins Glen, New York

Figure 130

Detail of early diagenetic leaching surfaces from figure 129.The fine dolomite crystals form grey clusters in the clear salt; directly below the leaching surfaces their concentration increases as a result of selective dissolution.Periods of leaching and channel-filling alternated repeatedly, the latter sometimes leaving chevronned turbulence track (bottom right).

Nicols not crossed, magnification 40 diameters

Upper Syracuse, Watkins Glen, New York

possibly caused by algae dragged at the bottom by the same currents. Deposition of carbonate in the algae may be the result of continued photosynthetic absorption of CO_2 before the plants died; rapid filling of the hollows by salt preserved their internal structure.

Nodular varieties of dolomitic halite are much more common. Most of these resemble the shaly or argillaceous-dolomitic halites, but the semi-continuous network separating the nodules, lenses and euhedral crystals of halite consists here of dolomite. The network sometimes joins dolomite laminae, covering salt pyramids which lie on the dolomite bed with their points projecting upwards (Figure 65). This may indicate that some crystals covered the dolomite before the material of the overlying dolomite network was deposited upon them.

Often the salt is predominant and only single relics of dolomite occur. The colour of the salt is generally much lighter than in the more argillaceous varieties. Some of the relics are vertically asymmetrical (Figure 52); their lower part fills pyramid-shaped depressions, while the upper surface is comparatively flat. The dolomite here is coarse dolomite silt, as in the Middle Syracuse Dolomite Suite, and as there, it also contains a few coarse quartz silt grains. Strongly bituminous dark laminae, halite lenses, and a few anhydrite needles also occur. In most cases, however, the relics are vertically symmetrical, their shape being the negative of the surrounding salt crystals. Irregular semi-continuous relics (Figure 131) join to form a pattern not expressive of any upward or downward orientation.

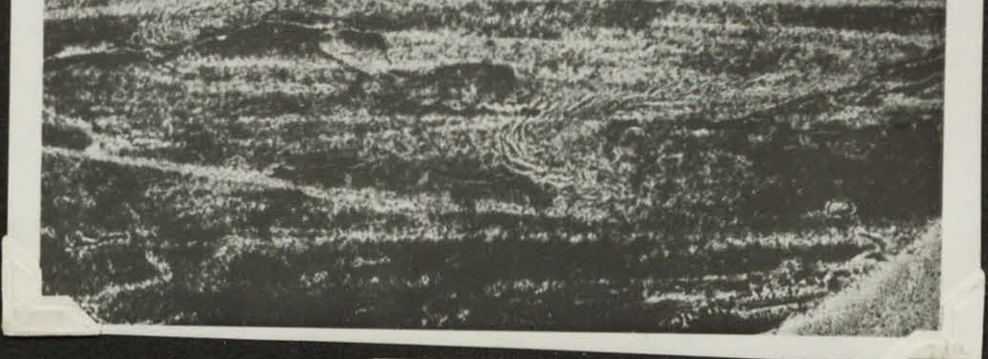


Fig. 129

Fig. 130

Figure 131

Loosely connected dolomite network in dark grey salt.

Actual size

Lower Syracuse, Watkins Glen, New York

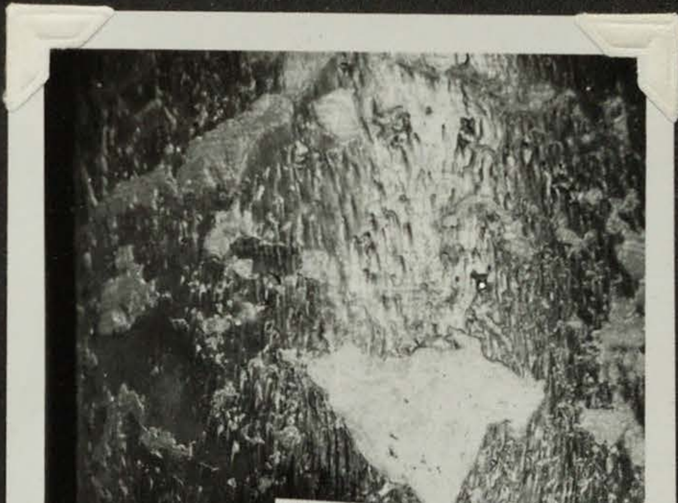


Fig. 131

Like the genesis of the non-laminar shale structures, that of the non-laminar dolomite structures is highly complex. It commenced with carbonate deposition over an uneven surface covered by loose salt crystals like those found in artificial evaporating pools. The carbonate mud precipitated partly biologically (algae), partly inorganically. Some of the silt-sized material may have been carried in by wind, blowing from the direction of carbonate dunes.

The original structure of these primary sediments became disrupted by the processes discussed above, particularly by early diagenetic subaerial and subaquatic leaching and recrystallization. The solution surfaces marked by the enrichment of dolomite grains indicate occasional dilution resulting in superficial leaching; at places the leaching may have been more extensive, widening the intricate channel system among the loosely packed crystals. These channels subsequently became filled with carbonate mud washed in from above. During periods of desiccation, repeated recrystallization of the salt in the capillary zone may have accentuated this non-laminar structure. Finally, late diagenetic flowage partially obscured the primary relationships.

*

*　　　*

The salt, whether dolomitic, anhydritic, or argillaceous-dolomitic-anhydritic, was apparently deposited in extensive sheets of very shallow water, extending northwards from the normal sea and periodically flooding the entire area of the "Salina Basin". This sheet of superhaline brine covered the non-halitic sediments, mainly argillaceous floodplain muds and algal biostromes. The high rate of evaporation repeatedly increased the inclination of the water table to the land, exposing large surfaces of salt to subaerial weathering. The salt, both subaerially exposed and the subaquatic, became repeatedly flooded by mud-carrying terrestrial waters coming from the peneplenized Taconic Mountains in the east. Such inflows, as well as occasional intrusions of undersaturated waters from the sea, resulted in intensive leaching of the intercrystalline channels between the loose salt crystals. These channels were subsequently filled by argillaceous or carbonate mud. Each subsequent exposure to air of such halite-carbonate-clay rocks led to increasing recrystallization of the salt in the capillary zone and to concentration of non-halitic impurities in an intercrystalline network.

The influence of seasonal changes in the rate of evaporation was less intensive in areas farther from the Taconic Mountains. They resulted in shifts of the concentration boundaries, replacing from time to time salt deposition with anhydrite deposition. Reflecting their respective concentrations in sea water, only fractions of a mm of anhydrite were deposited during the same span of time which deposited several cms of salt. Thicker anhydrite beds (up to 10 cm thick in the Upper Syracuse salt in Cleveland) indicate that lower concentrations, probably reflecting lower rates of evaporation, became occasionally established over the area for several centuries. Due to such changes, it is difficult to calculate depositional time for the entire sequence by counting the number of salt and anhydrite laminae.

Pattern of cyclicity

Statistical considerations

In general, then, though transitional facies are common, three major lithologies occur in the area:

Halite
Dolomite, most of it biogenic or bioclastic
Shale

Though the shales overlying dolomite or halite are frequently lined at their base with and contain lenses of anhydrite, normally Ca-sulphate does not form thicker layers. Disregarding transitional varieties and minor interbeds, the following Salina sequence of the axial zone (at Watkins Glen) provides material for some statistical analysis:

Camillus Formation		S
Upper Syracuse,	F4	H
		S
		H
	F3/4	D
		S
		H
		D
	F3	H
		D
		H
	F2/3*	D
		H
		S

/S=Shale
D=Dolomite
H=Halite/

*Incomplete

Upper Syracuse,	F2/3	D
	F2	H
		D
		H
	F1/2	D
		H
		S
		D
		S
	F1-2	H
		S
		H
	F1-2/1	D
		S
	F1-1	H
	E4	S
		H
		S
		D
		S
	E3	D
		S
		D
		S
	E2	H
		D
		S

Middle Syracuse,	E1	H
		D
		S
		D
		S
		D
Lower Syracuse,	D3	H
	D3/2	S
		H
		S
		H
		S
		H
		S
	D2	H
	D2/1	D
		S
		H
		S
	D1	H
Upper Vernon	C	S

In the above presentation, 62 layers are differentiated. Of the 61 contacts, shale overlies halite in 15, underlies it in 12 cases; shale overlies dolomite in 7, underlies it in 10 cases; halite overlies dolomite in 10, underlies it in 7 cases.

Thus dolomite occurs above shale 43% more frequently than either below it or above halite. Halite occurs above shale 20% more often than above dolomite, but both frequencies are considerably higher (by 71 and 43 %, respectively) than that of halite underlying dolomite. Finally, shale overlies halite more than twice (2.14x) more often than it does dolomite. An SDHSDH succession is, then, preferred.

The actual preference of such a succession is even more evident if we observe the probability ratios of such triplets which contain all three major lithologies. Six such triplets are possible:

SDH	HDS
DHS	DSH
HSD	SHD

which can be grouped into two infinite series:

SDH, DHS and HSD form.SDHSDH...

HDS, DSH and SHD form.HDSHDS...

These triplets (with each layer participating in three triplets as their initial, middle and final member, respectively) occur in the succession the following times:

SDH	5	HDS	2
DHS	6	DSH	3
HSD	6	SHD	3

The succession, then, contains more than twice (2.12x) as many SDH than HDS-type triplets, rendering the trend even more evident.

As far as triplets of alternating type are concerned (whose initial and final members are the same), their distribution follows the following pattern:

SHS 9	HSH 8
DHD 4	HDH 5
SDS 5	DSD 4

So halite alternates with shale almost twice (1.9x) more frequently than with dolomite, while the latter alternates with equal frequency with halite or with shale.

This lays the basis for the selection of the most frequent (modal, Duff and Walton,1962) triplets. At Watkins Glen, out of 60 triplets, 28-28 % belong to triplets fitting into the SDHSDH and SHSH series, while the frequency of the other possible series (HDSHDS, HDHD and SDSD) ranges only from 13-15 %.

So while the three major rock varieties can succeed each other in any order, the distribution of frequencies is far from random: triplets of SDH-type and alternations of SH-type occur about twice as frequently as any other comparable set. These two modal cycles replace one another horizontally as well as vertically, as dolomite appears and increases in thickness between the shale and the overlying halite. The thickness of this dolomite appears to be inversely proportional to the detritus supply, often missing in the most detritic periods and areas, and becoming increasingly significant in the less argillaceous sections of the Middle and Upper Syracuse, particularly farther from the detritus source area. Both in the more and in the less argillaceous areas and sections, the occasional occurrence of anhydrite at the bottom of the shale of both modal cycles supplements them to form composite cycles of ASDH and ASH composition.

Genetical relationships

The Taconic mountain-islands, marking the eastern boundary of the Appalachian Basin, are bordered to the northwest by red and green siltstones with interbedded sandstones. Farther to the west-northwest, these grade into the thinner grey and green shales exposed in the axial zone,at Watkins Glen,(Fig. 132), alternating with thick layers of considerably argillaceous halite and thinner, also argillaceous dolomite. Further to the west, towards the platform zone, the thickness of the shales decreases, while that of the less and less argillaceous halites and dolomites increases; many of the thinner beds and laminae of anhydritic shale, particularly where alternating with halite, grade into more or less clay-free anhydrite.

Toward the open sea in the south (Fig. 133), the dolomites pass into limestones, containing increasing amounts of bioherms built particularly by stromatoporoids and corals. The occasional presence of the same colonies in the dolomites of the north suggests that originally they, too, consisted of Ca-carbonates (aragonite and Mg-calcite), and replacement of the calcium carbonates by dolomite has been due to the increasing Mg-concentration of sea water concomitant with evaporite formation.

Signs of temporary desiccation are common in all of the major sediment types. They are most evident in the shales, where they are reflected in the abundance of finely laminated clay chips, anhydrite

<u>Figure 132</u>

Lithologic changes during the progressive phase of an evaporite-bearing cycle across the northern Appalachian basin. Transition from clastics to evaporite is concomitant with decreasing terrestrial inflow.

1-Fluviatile red beds; 2-Variegated siltstones; 3-dolomitic shales; 4-Calcareous shales; 5-Argillaceous dolomite; 6-Dolomites, partly bioconstructed; 7-Limestones, partly bioconstructed; 8-halite; 8-$CaSO_4$-content significant.

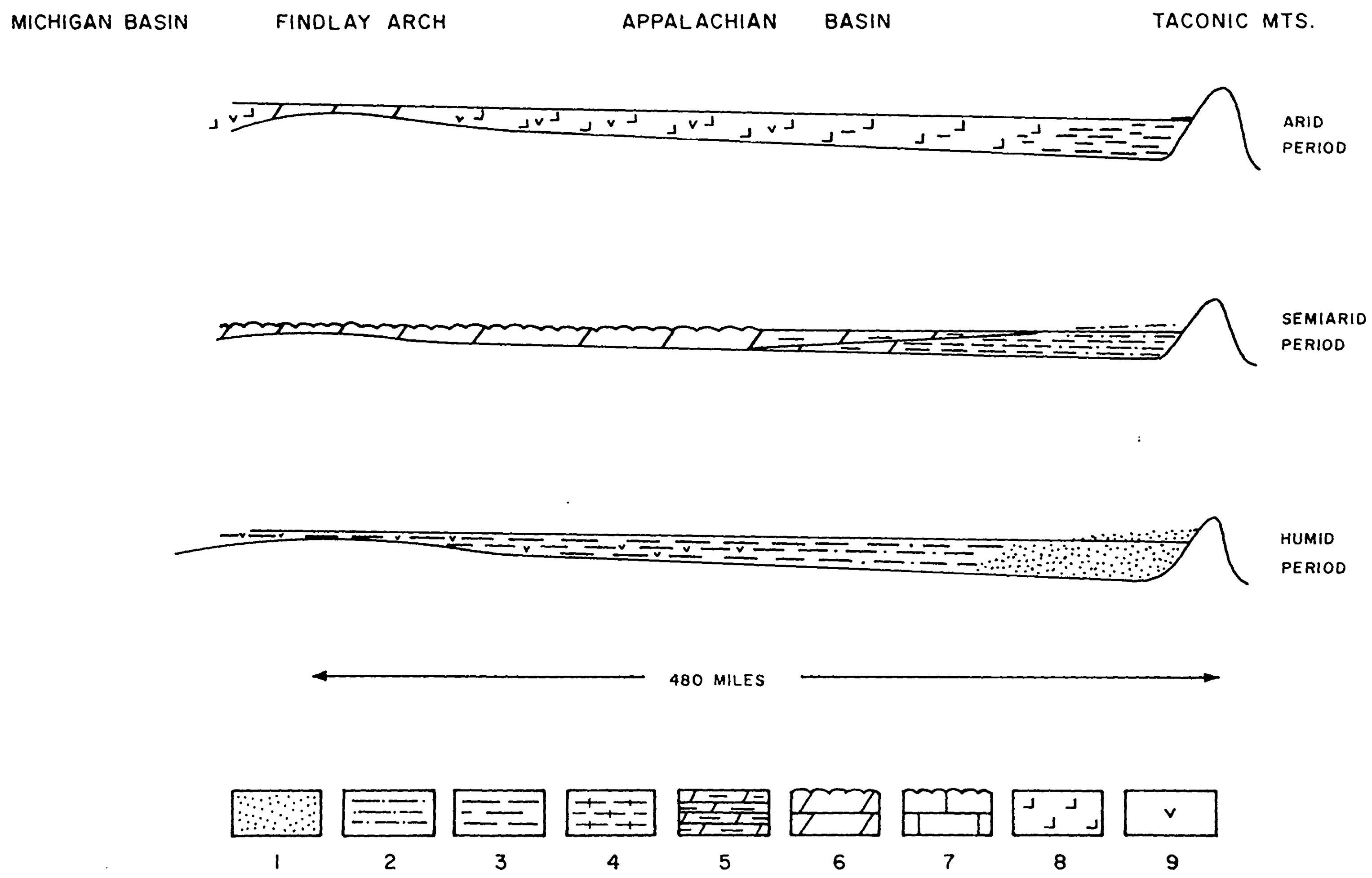
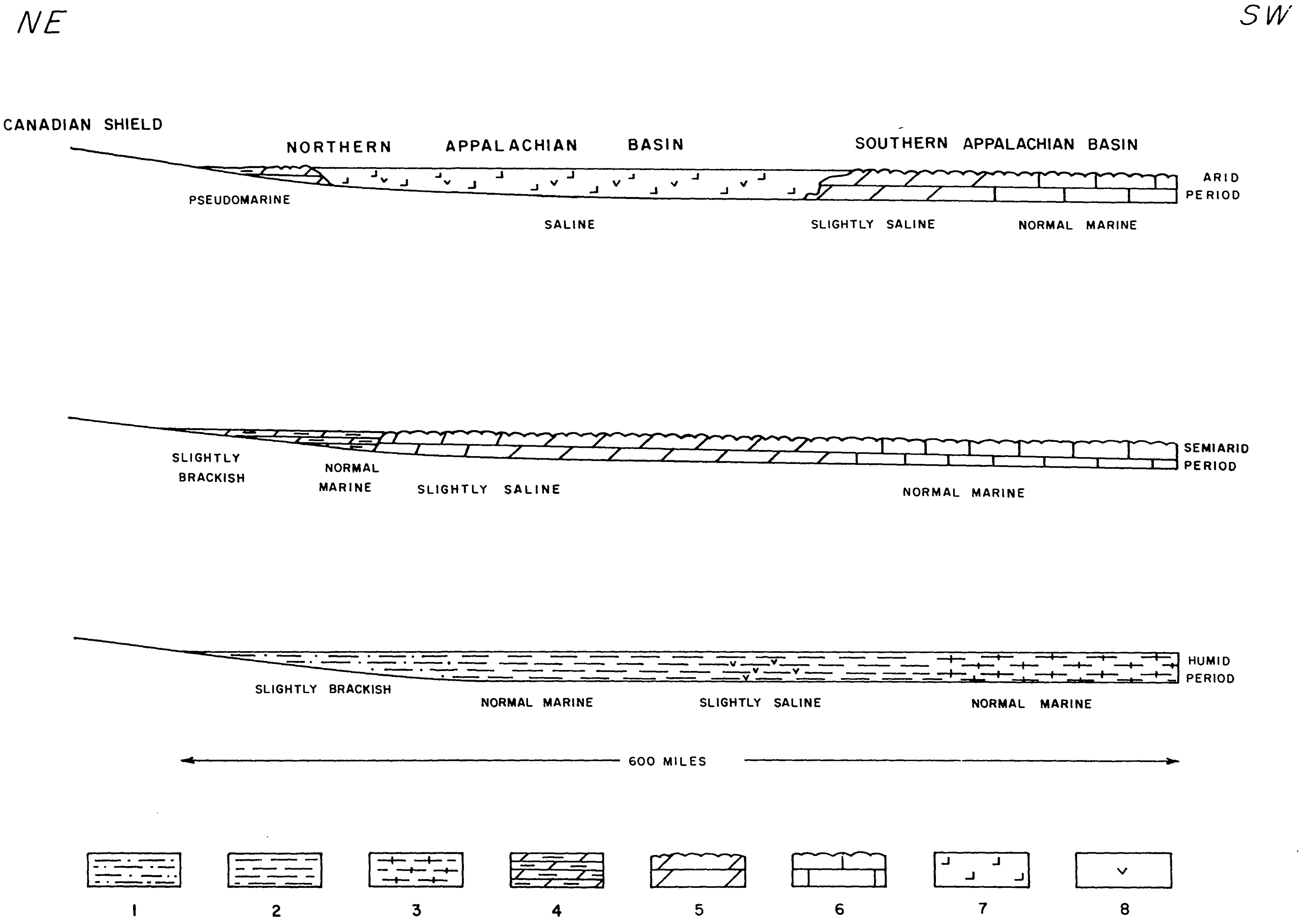

MICHIGAN BASIN
FINDLAY ARCH
APPALACHIAN BASIN
TACONIC MTS.
ARID PERIOD
SEMIARID PERIOD
HUMID PERIOD
480 MILES
1
2
3
4
5
6
7
8
9

Fig. 132

Fig. 133

Lithologic changes during the progressive phase of an evaporite-bearing cycle along the axis of the Appalachian Basin. Transition from clastics to evaporites is concomitant with decreasing terrestrial inflow.

1-Variegated siltstones; 2-Dolomitic shales; 3-Calcareous shales; 4-Argillaceous dolomite; 5-Algal stromatolites and reef dolomite; 6-Limestone; 7-Halite; 8-$CaSO_4$-content significant.


NE
SW
CANADIAN SHIELD
NORTHERN APPALACHIAN BASIN
SOUTHERN APPALACHIAN BASIN
ARID PERIOD
PSEUDOMARINE
SALINE
SLIGHTLY SALINE
NORMAL MARINE
SEMIARID PERIOD
SLIGHTLY BRACKISH
NORMAL MARINE
SLIGHTLY SALINE
NORMAL MARINE
HUMID PERIOD
SLIGHTLY BRACKISH
NORMAL MARINE
SLIGHTLY SALINE
NORMAL MARINE
600 MILES
1
2
3
4
5
6
7
8

lenses (apparently formed in the capillary zone), and polygonal mudcrack systems. Discontinuous shale chips, and mudcrack systems restricted to single laminae, characterize also the basal, anhydritic zones of shale layers.

In the dolomite, desiccational phenomena occur less often. They are manifested in complex capillary channel systems filled with salt, halite spherules and anhydrite needles occurring in distinct horizons, and desiccational cracks at the top of the dolomite layers. The latter are often covered by unsorted, unrounded dolomite detritus.

In the halite, desiccation can be postulated indirectly from desiccational phenomena displayed by intercalating shale laminae. Alterations at the top of thicker salt layers, including extensive recrystallization, destruction of the original laminar structure, appearance of channels filled with anhydrite detritus, fissures filled with clay, and fragmented shale laminae surrounded by partially recrystallized haematitic salt also reflect temporary exposure to air succeeded by, or alternating with, flooding by fresh water or by lagoon brines.

Apparently, then, all three major sediment types were subjected to occasional desiccation, and although this seems to have been less frequent in the chemical deposits, the depth of the water cover does not appear to have been the only major factor determining the sedimentary facies.

A major factor controlling deposition was, apparently, the intermittent supply of detritus and fresh water from the land areas. Though these land areas included a few temporarily emerged portions of the Canadian Shield (as witnessed by the presence of red and green shale alternating with halite along the northern margin of the Michigan Basin), they consisted primarily of eroded remnants of the Taconic Mountains, forming long islands along the eastern margin of the submerged continent. They provided detritus and fresh water inflow into the basin, forming the only barrier in the path of the winds. The fact that these islands then trended parallel to the westerlies probably contributed to the arid climate.

In such an arid climate, even short periods of heavy rainfalls, whether falling over land or sea, could have profound effects on a shallow sheet of brine. (The traces of such rainfalls are preserved in abundant raindrop marks on the bedding planes of desiccated shale layers). Apart from partially eroding and leaching sediments exposed to the air, the inflowing fresh water could inhibit evaporite deposition by maintaining a surface sheet of undersaturated water. Analysis of vertical and horizontal trends in the bromine content of the Salina salt led Kunasz (1970) to believe that " the mechanism causing (the) dilution (responsible for the deposition of intercalating anhydrite laminae) appears to have been meteoric water which dissolved halite exposed during the dry 'season' and recycled it to the basin". Equally important is the fact that, by sweeping in large amounts of clay from the land area, the rains covered the previously deposited evaporites with a protective sheet of mud.

Such rapid inflow of mud raises the sediment level in the basin. Upon the return of aridity, much of this sediment would be exposed to air even if it was slightly above sea level, since the inflow of sea water, restricted by friction, could not entirely compensate for evaporation over such vast shallow expanses and would form a surface tilted landward. Only occasional storms and tidal waves would flood the sediments, bringing in marine plankton and saturating the capillary brines and surface pools. Subsequent evaporation would then produce surface crusts and subsurface nodules of Ca-sulphate or of halite coated with Ca-sulphate. Desicational and pressure cracks would be filled with similar lateral sequences of carbonate , sulphate and chloride.

Decrease in precipitation and consequent decrease in sediment supply would leave the subsidence uncompensated, thus preparing the way to the return of the sea. If the sea of normal concentration is far, pre-saturation areas are large, and bars extending from small micro-deltas of the island rivers are still effective in limiting sea water inflow, then the transgressive sea will be hypersaline depositing salt over much of the area. This was particularly common during Lower and Upper Syracuse times, resulting in alternating layers of halite and mudstone. If, however, the ingressing part of the sea is relatively undersaturated and becomes even more so upon occasional dilution by terrestrial waters, as was the case particularly during early Middle Syracuse times, extensive areas become favourable for the growth of CO_3-selecting organisms. Accordingly, extensive sheets of biostromes develop (mainly algae and stromatoporoids, with more bryozoa and Tabulata corals toward the open sea), reaching from the normal sea into areas often characterized by halite deposition. Increasing distance from the detritus source and relatively low salinity (often concomitant with semi-arid climate) are favourable for the growth of such biostromes (Figure 132).

Much of the biostromes and associated carbonate sand and silt bars reached to sea level and was subject to frequent desiccation, partic..ularly after their growth had stopped. The resulting precipitation of halite and Ca-sulphate in the capillary zone, in the form of spherules, concretions and cavity fillings, not only decreased the permeability of the rock, but also shifted the Ca/Mg ratio in the interstitial brines, thus increasingly dolomitizing the biostromes.

The growth of the biostromes was self-inhibitive: increasing the friction and decreasing the front of the ingressing sea water, they extended the area of pre-saturating pools, ultimately producing the same degree of isolation which had earlier resulted from the abundant inflow of clastics. Thus the biostromes gradually receded, partly toward the open sea and partly to the shoreline where terrestrial inflow maintained pseudomarine conditions. Between these two zones of normal sea concentration, and eventually in the pseudomarine zone as well, the biostromes became gradually covered by highly saturated brines, depositing Ca-sulphate bearing halite and, during rainy seasons or periods, argillaceous Ca-sulphates or Ca-sulphate bearing muds, depending on the distance from the detritus source area. The rapid deposition of halite, just like the rapid inflow of mud or the rapid growth of biostromes, resulted in occasional withdrawal of the sea from the rapidly rising sediment level, leading to deepreaching alterations in the subaerially exposed salt.

S Y N O P S I S

On the preceding pages, we have reviewed two evaporite-bearing sequences, distant from one another in both time and space. Both sequences consisted of the same lithologies (shale, dolomite, anhydrite and halite) and both displayed intricate cyclic structures. There were many differences, too, like the absence of potassium-bearing salt horizons and the subordinate presence of anhydrite in the Salina Group of the Appalachian Basin. Many of the similarities can be accounted for by noting the essentially similar tectonic-paleogeographic position of the two basins.

Tectonic-paleogeographic setting

The deposition of either sequence took place in an elongated basin, located between a parallel mountain range and a stable platform, and extended considerably over the platform. The basin served, through most of its history, as a collector of debris swept in from the rising mountain ranges of the adjacent orogenic zone. Its fill, several thousand meters thick, consists mainly of detritic suites and represents a time span of era magnitude. Thus the fill of North German Basin represents a time span extending over most of the Upper Paleozoic and Mesozoic, that of the Appalachian Basin was accumulated through most of the Paleozoic. The depositional time of the evaporite-bearing sequence formed only an infinitesimally small portion of the depositional history of the basin, though the sequence itself comprises a more significant portion of the total fill. It was deposited during the later stage of a period of relative tectonic calm coinciding with a period of aridity on the basinward side of the mountains. The lowered relief and the dry climate eliminating much of the detritus inflow, the subsidence remained uncompensated, permitting a marine inundation and subsequent mostly chemical deposition.

The inflow of sea water was, at the same time, sufficiently inhibited (possibly by a bioconstructed bar over submerged Caledonic remnants in the Zechstein and by submerged deltas in the Appalachian Basin) to compensate only imperfectly for the high rates of evaporation in the inundated arid areas, thus causing rising brine concentrations and frequent deposition of even the more soluble salts. The inhibition of inflow was apparently more complete in the Zechstein than in the Appalachian Basin, permitting in the former recurring deposition of the most soluble K and Mg salts. The terrestrial inflow was, on the other hand, more significant in the Appalachian Basin, depositing shales in more significant proportions. Increased bacterial activity in the latter basin, possibly related to the lower concentrations and to the wider spread of algal life, together with the higher Fe concentrations corresponding to the higher levels of terrestrial inflow, may have led to synsedimentary reduction of the precipitating $CaSO_4$ and ultimately to increased pyrite deposition and a relative $CaSO_4$-deficiency. Similar processes have been observed by Neev and Emery (1967) in the Dead Sea.

Sedimentary imbalance

In such a milieu, alternating deposition of diverse lithologies was inevitable. Even slight changes in the temperature were sufficient to make one salt stable and the other unstable in solution. Even slight improvements in the communication with the open sea could permit the inflow of large volumes of sea water, altering the concentration pattern of the basin and redissolving large portions of its sediments. Even slight increases in the rate of precipitation over the adjoining mountain ranges could produce massive inflows of fresh water, changing the concentration and temperature equilibria in the basin and sweeping in large amounts of debris accumulated during the preceding dry period. The occasional filling of the basin by rapidly depositing salts or muds as well as the exclusion or expulsion of sea water by temporary topographic barriers or offshore winds could expose large areas of the basin floor to subaerial evaporation, further increasing the concentration of brines retained in the interstices and effecting deep-reaching early diagenetic changes. In such a sensitive system, periodic changes in the conditions of deposition were readily reflected in sedimentary cycles.

Annual cycles

The most distinct non-diurnal periods on the surface of this planet are produced by the alternation of winter and summer, rainy and dry season. Annual cycles are obseervable in recent evaporite deposits, and the ancient ones present good analogues. Perhaps the finely laminated basal carbonates and sulphates of the lower Zechstein cycles, studied in great detail by Richter-Bernburg (1957-) are the best examples. In these cycles, organic-rich clay and carbonate, or, in the sulphates, carbonate (lined by an organic-rich film) and sulphate lamellae alternate. The annual nature of most of these cycles is generally accepted. Though observations by Neev in Dead Sea sediments indicate that corresponding varves do not form every year at present, they appear to have done so in the past (Neev and Emery 1967). In the Zechstein, Braitsch (1962) tends to view them, if they are annual, as the products of seasonal concentrational shifts in the surface waters, toward and away from the open sea. The analogues he presents do not, however, appear to be of the required magnitude; seasonal inflow of terrestrial waters seems to provide a more adequate mechanism of control.

The more controversial cycles involving halite are much more widespread throughout the sequence. Halite layers with parallel laminae of anhydrite or shale are common in both the Zechstein and the Salina Groups. Historically, they have been considered annual "Jahresringe" (Zimmermann, 1915). At first, they were attributed to changes in the temperature: anhydrite, less soluble in hot than in cold water, would precipitate in the summer; halite, less soluble in cold water, would be deposited during the winter. Braitsch (1962/72) points out, however, that in brines saturated in NaCl, $CaSO_4$ forms only 0.7% of the total $NaCl+CaSO_4$ content, compared to the 4-5, rarely up to 9 per cent observed in the laminated halites, therefore temperature fluctuations alone cannot account for the rhythmic structure of the latter. Bromine studies indicated that the anhydrite laminae were deposited from diluted brines, so that only the source of the dilutant had to be established. Several mechanisms of periodic marine inflow have been suggested, ranging from the original hypothesis of Zimmermann (1915) involving a bar breached annually by the Monsoon and re-formed by the counter-Monsoon, to the mechanism of seasonally shifting concentrations in the surface brines , suggested by Braitsch (1962/71). The first interpretation requires somewhat arbitrary conditions to be sustained for much of the depositional time, the second would explain the lamination only within a narrow transitional zone. Periodic inflow of rain water and terrestrial run-off (Stewart, 1963) seems again to offer a more adequate mechanism.

Observations in the Salina Group of the Appalachian Basin may perhaps be taken as direct evidence in favor of the control of halite lamination by terrestrial run-off. Anhydrite laminae appearing at regular intervals in some of the salt beds of the platform zone become progressively more argillaceous in the direction of the mountains and are replaced by anhydritic shale laminae in the proximity of the orogenic zone. In such a model, the rain water of terrestrial run-off may also provide the Ca-ions necessary for the precipitation of $CaSO_4$, at the same time freeing the Mg balancing the SO_4 in the brines to effect the Mg-metasomatism of the carbonates.

The thickness of annual halite cycles has been an important consideration in interpreting the genetic conditions of evaporite-bearing sequences. If these cycles, several cm thick, are annual precipitates, than the depositional rate of the evaporite-bearing sequence greatly exceeded the rate of subsidence, so that the evaporites had to be deposited in a pre-existing, and initially deep, basin. Taking advantage of a rare opportunity of comparing the number of laminae in halite and in its faciological anhydrite equivalent, Jung (1959, 1968) found that the latter contains approximately ten times more annual lamellae. His subsequent studies revealed the existence of the required number of anhydrite laminae in the halite as well, but only about 1/10 of these were visible in the outcrop, and the rest could be seen in transmitted light only. This observation was held by him to signify that the truly annual laminae of halite are ten times less thick than they had been thought before, and consequently their rate of deposition does not significantly exceed the possible rates of subsidence. This would waive the necessity of postulating a pre-existing deep basin for the deposition of halite and permit the application of a shallow-water model; for such far-reaching conclusions, however, more evidence is yet needed.

Solar cycles

If the question of the relatively simple annual cycles is so controversial, that of the solar cycles is naturally much more so. Baur (1961, in Braitsch, 1962) found that even as far as the general weather pattern is concerned, the 11-year period is well within the range of chance, though Braitsch notes that this may be due to the short duration of systematic measurements (extending over a few solar cycles). Baur finds that "in the rhythmic analysis of certain weather elements an 11-year cycle can be seen, but the amplitudes are always so small that they in no way explain weather anomalies." Whether they explain sediment anomalies, is still more controversial. Fundamental studies by Richter-Bernburg (1957-1963) in the laminated basal anhydrites of the Stassfurt Cycle revealed recurring anomalous lamina thicknesses at 11-varve periods. Longer periods (22,34, 45; 95, 180-200 and 380-400 varves) were also observed. Critics of this approach, particularly Braitsch (1962) and Duff et al. (1967) point to the subjective judgment involved in selecting the varves of anomalous thickness, and the lack of mathematical analysis in determining the periodicity. Systematic harmonic analysis would do much to clarify the problem.

In the laminated halites, a periodicity close to the length of the solar cycle might be seen in the findings of Jung (1957, 1968) discussed above, as well as in the findings of Roth (1970, oral comm.) revealing a fine alternation of hematitic-sylvitic and pure halite laminae between two laminae of argillaceous anhydrite. In these cases, the sharp difference between what is thought to be an annual and what is considered an 11-yearly lamina seems to be incongruous with the gradual nature of the solar cycles; one might theorize that sediments are not affected until certain boundary conditions are passed and then the change is sharp, but as yet there is no observational evidence in favor of such a hypothesis.

Intermediate and major cycles

Beyond the minor cycles discussed above, both evaporite-bearing sequences display cycles of longer periods. Some of these are the major cycles of the given evaporite-bearing sequence (defined arbitrarily as the smallest number of cycles into which the sequence can be consistently divided); the others are a few meters or tens of meters thick and provide the internal structure of the major cycles.

Some of these cycles are symmetrical, in that their upper portion approximates the mirror image of the lower one; others are asymmetrical cycles displaying a progressive or a recessive pattern. In the progressive cycles, the solubility of the sediments increases upwards, and the most soluble sediments at the top of one cycle are overlain, either abruptly or with a thin transitional zone, by the least soluble basal sediments of the following cycle. In recessive cycles, this pattern is inverse: the solubility of the sediments decreases gradually and rises abruptly.

Though these groups are morphological categories, it has been attempted to define their genetic content. The pattern originally envisaged for the major cycles was that of a large body of sea water, cut off from the sea and gradually evaporating. This brine would then deposit a series of increasingly soluble salts until its complete desiccation.

Quantitative considerations indicated that a single evaporating inland sea could not have deposited the salts of a major cycle: to account for the salts of the Stassfurt Cycle, a sea about 50,000 meters deep would have had to evaporate. Therefore it has been assumed that the separation from the open sea was only partial, and an influx of sea water, continuous or periodical, replaced the evaporating water. Each major inundation would result in an abrupt increase in the dissolving capacity of the brines and partial dissolution of of previously deposited salts. The barrier would then be gradually restored and a new cycle of progressively soluble sediments would be deposited. Minor reversals in this trend (e.g. from NaCl to $CaSO_4$) would reflect minor improvements in communication with the open sea; major reversals (e.g. from chlorides to shale and dolomite) would reflect a major inundation.

This model required further modification. Thus, it has been found that a one-way flow of sea water into the basin would not account for the observed sediment ratios: in the Zechstein, the amount of carbonates and $CaSO_4$ is considerably more than what would correspond to the amount of the more soluble salts (Braitsch, 1962). Therefore a counter-current or reflux of heavy brines had to be postulated, egressing into the open sea below the level of the incoming surface waters. Only in the late stages of chloride deposition would this reflux cease.

The concept of reflux made the conditions of evaporite deposition more tenuous. Complete separation of a body of sea water from the open sea would, under arid conditions, necessarily result in the deposition of progressively more soluble salts; if the separation is incomplete but sea water flows only toward the evaporating basin, this process would take more time but it would produce a similar (and much thicker) sequence of salts. But if the flow is in both directions, no significant increase in concentrations has to ensue. Thus the Mediterranean, located under an arid climate and fed by a strong influx of oceanic water, shows only a minute increase in salt concentration, in spite of the narrowness of the Straits and the height of the submarine threshold inhibiting the reflux.

If such a narrow and shallow entrance is hardly sufficient even for a minor increase in concentration, then either a bar has to be postulated, reaching almost to sea level most of the time, or an extensive saturation shelf has to be envisaged. It is questionable if dynamic friction in the latter would inhibit the reflux sufficiently to permit the required rise of concentrations; whereas the former would eliminate reflux unless the bar consisted of a permeable substance. Therefore a bar of porous sediments (reefs?) topping submerged Caledonic remnants is envisaged for the Zechstein, and an extensive saturation shelf with or without bars is suggested for the Appalachian Basin. In this context, we shall review the factors exercising control over the intermediate and major cycles.

1. Factors affecting the bar area

Due to the crucial influence of the bar in regulating the balance of influx and reflux, any factor affecting it would also affect the entire area of deposition. Tectonically, we have assumed that the bar of the Zechstein Basin was related to submerged remnants of the Caledonian Mountains. During a period that was quiet even in the active Variscan orogenic zone, the eroded-submerged older mountains probably did not exhibit much tectonic activity, but even slight intermittent subsidence would have been enough to lower the bar and breach a barrier reef, letting in large volumes of sea water into the basin. Subsequent sedimentation, and the growth of reef-building organisms, would gradually seal the bar again. Such intermittent breach and gradual growth of the barrier would result in the formation of progressive cycles in the basin.

An eustatic rise of sea level would have the same effects, if it were followed by an upward growth of the bar (Stewart, 1963). Duff et al. (1967) see in the progressive pattern of the cycles an indication that changes in sea level may indeed have been the controlling factor. Glacially induced eustatic changes during the late Permian have generally been considered likely, though Meyerhof (1971) believes that the glaciation did not coincide with the deposition of the Zechstein salt sequence.

If the crucial top portion of the bar is bioconstructed or consists of soft sediments, any factor affecting sedimentation would affect its height. A climatic control could be exercised through increased deposition of the less soluble salts along the inner side of the bar during the more arid periods. These precipitates would plug the pores of the bar sediments, thus inhibiting the reflux of brines and increasing the concentration of brines in the basin. The higher concentrations would be reflected in the reflux as well, leading to further plugging. The return of the less arid climate would reverse the process; the inflow of terrestrial waters would produce a surface sheet of low concentrations, inhibiting the deposition of salts in the bar area as well, and thus increasing the permeability of the newly formed sediments.

A self-regulating sedimentary control can also be envisaged. As the concentration of brines increases in the basin, the concentration of the reflux also rises. This concentrated reflux would first inhibit, then stop the growth of the reefs crowning the bar. Continuing subsidence, uncompensated by reef growth, would lower the bar level and open up the basin to the influx of marine waters. The brines and consequently the reflux would be diluted, thus reestablishing conditions favorable for reef growth in the bar area. Renewed separation of the basin would again raise the concentration of its brines, starting a new cycle of evaporite deposition.

2. Factors affecting the basin

Paradoxically, the depositional cycles of an evaporite-bearing basin are probably least influenced by controls affecting the basin itself. Periodic increases in the rate of subsidence may result in repeated inundation of the coastal plains, producing cycles of subtidal-supratidal deposits. Local differences in the rate of subsidence may create deeps favorable for the deposition of K-bearing salts. Periodic rises of humidity may decrease evaporation and thereby inhibit the precipitation of salts, though usually the effect of the inflowing terrestrial run-off would by greater than the direct influence of risen humidity. The effect of rains is greatest if the basin is essentially filled with salts and entirely cut off from the open sea; in such cases leaching of the exposed sediments by rain is followed by the deposition of pure salts comcomitant with the evaporation of the water introduced. The recessive cycles thus formed usually display signs indicating horizontal redistribution, though actual current bedding occurs rarely (Swath Salt).

3. Factors affecting the orogenic zone

As attention in the past has been focused upon controls influencing the hypothetical bar area, it might be promising to review the factors affecting the orogenic zone. Even during such a period of tectonic calm, minor movements may take place increasing the gradient and thereby the inflow of detritus into the basin. The increasing amount of shales, appearing even between salt layers during the third and fourth Zechstein cycles, may reflect the start of tectonic activities culminating in the Pfalzic movements. Indirectly, tectonically induced rise of the mountains may also increase precipitation over them and consequently increase the inflow of terrestrial water into the basin.

This leads us to the discussion of climatic factors. Periodic changes in rainfall over the adjacent mountains has a strong influence over the depositional processes of the evaporating basin. This is particularly true of the Paleozoic, when the balancing effect of terrestrial vegetation was still limited. The debris accumulated during the preceding dry period was the first to be carried into the basin, giving rise to upward fining cycles, further modulated by the shifting channels of terrestrial run-off. Pollen studies carried out on some of the shales at the base of the major Zechstein cycles (Kosmahl, 1969) indicate that these indeed formed under a more humid climate than the salts, though the increase in humidity did not have to extend into the basin itself. In the Appalachian Basin, the progressive cycles seem to be due to similar climatic changes in the orogenic region. Apparently, periodically increasing rainfall in the source area resulted in flooding and silting up of the basin by muds; with subsequent decrease in the rainfall over the mountains, these muds were covered by carbonates deposited in a pseudomarine belt receding toward the mountains. Ultimately, the basin was flooded by a hypersaline brine depositing salt. Both in the Zechstein and the Appalachian Basins, the thickness and grain-size of the shales increases toward the mountainous source area.

Massive development of carbonates and sulphates along the basin margin during the lower Zechstein cycles may also be due to the inflow of $Ca(HCO_3)$-bearing terrestrial waters into brines rich in Mg and SO_4 ions. Much of the SO_4 would be deposited in the form of $CaSO_4$, and the Mg would effect dolomitization and magnesitization of the accompanying carbonates. Though Braitsch (1962) considers this process unlikely for the Stassfurt Cycle as it would deplete the brines in $MgSO_4$ and inhibit the subsequent deposition of $MgSO_4$-bearing salts in the Stassfurt Potash Zone, one may not concur with this opinion: as the ratio between the anhydrite and potash salts of the Stassfurt Cycle does not correspond to the ratio of their composing ions in sea water, the brines providing the material for the deposition of $MgSO_4$-bearing salts did not have to be the same ones from which the marginal $CaSO_4$-s were deposited.

When reviewing the potentially great significance of climatic control, affecting the evaporating basin through its in fluence over the orogenic zone, a word of caution might be required. The periods of alternately higher and lower precipitation exercising this control do not have to be due to global climatic changes. Such global fluctuations of the climate do exist; they are caused by the cumulative effect of the nutation of the Earth's axis, precession of the equinoxes, and periodic changes in the eccentricity of the Earth's orbit. Though the irradiational changes caused thereby can be reliably calculated for much of the Pleistocene, further extrapolation cannot be done with sufficient accuracy to predict the interferences and determine the actual periodicity of the resultant curve.

But the influence of paleogeographic factors can be greater than that of the astronomical ones. The rise and erosion of mountains, the forming and silting up of aquifers, can change the direction, altitude and humidity of the winds sufficiently to alter the pattern of precipitation. The effects of sediment accumulation can also modify the pattern of terrestrial run-off; one may be reminded of the effects the diversion of the Amu- and Syr-Darya (the Oxus and Axartes of the ancients) from the Caspian into the Aral Sea in historic times had upon the concentration and sedimentation pattern of both aquifers.

E P I L O G U E

Perhaps because their physico-chemical behavior is better known than that of most other sediments, the depositional conditions of the evaporites are still problematic. Their delicate solution equilibria are easily disturbed, resulting in the formation of cycles of diverse thickness and origin. The periodic inflow of sea and rain water exercises the greatest control over these cycles: most cyclic phenomena attributed to one of these factors can equally be explained by invoking the other. The author feels that in the past less attention has been paid to the influence of rain water and terrestrial inflow than its significance would warrant. He conceives of the entire depositional area of an evaporite-bearing sequence as part of the boundary zone between the terrestrial and marine regimes, in which both regimes exercise their fluctuating influence.

To the human mind, striving to find the single Cause, models involving a common control for terrestrial and marine inflow might be preferable. Such models may include correlated periodic rise in the mountains and subsidence in the bar area, followed by erosion of the mountains and sedimentary growth of the bar; periods of increased rainfall resulting from an about 25% increase in the amount of water evaporated by the basin after its brines have been diluted by marine influx from concentrations corresponding to NaCl deposition to those of the sea water; and inhibition of plugging the pores of the bar sediments by the periodic inflow of terrestrial water.

The fact that the deposition of massive shales mostly conincides with that of the least soluble chemical sediments may justify such a unifying approach.

But it is possible that our desire for unity cannot be gratified on this level, that sea and rain assume control independently, some carbonates and calcium sulphates being the product of the inflow of Ca-bearing terrestrial water, and others depositing from the sea without any terrestrial contribution.

The calendar given us in the cyclic sequence is still waiting for its Champolion to solve its hieroglyphs.

References

Alling, H.L., 1928. The geology and origin of the Silurian salt of New York State: New York State Mus. Bull.275, 139 p.

------ and Briggs, L.I., Jr., 1961. Stratigraphy of Upper Silurian Cayugan evaporites: Am. Assoc. Petrol. Geol. Bull. v. 45, p. 515-547.

Baar, C.A. and Kühn, R., 1962. Der Werdegang der Kalisalzlagerstätten am Oberrhein: Neues Jahrb. Mineralogie Abh., v. 97, p. 289-336.

Berry, W.B.N., 1968. Ordovician paleogeography of New England and adjacent areas based on graptolites, in Studies of Appalachian geology - Northern and Maritime: New York: Interscience, p. 23-34.

------ and Boucot, A.J., 1967. Continental stability - A Silurian point of view: Jour. Geophys. Research, v. 72, p. 2254-2256.

------ and ------ 1970. Correlation of North American Silurian rocks: Geol. Soc. Amer. Spec. Pap. 102, 289 p.

Borchert, H., 1940. Die Salzlagerstätten des deutschen Zechsteins: Archiv für Lagerst. Forsch. H67:Berlin, 196 p.
------ 1959. Ozeane Salzlagerstätten: Berlin: Bornträger, 237 p.

------ and Muir, R.O., 1964. Salt deposits - the origin, metamorphism, and deformation of evaporites: London: Van Nostrand, 338 p.

Boucot, A.J., 1968. Silurian and Devonian of the Northern Appalachians, in Studies of Appalachian geology - Northern and Maritime: New York: Interscience, p. 83-94.

Braitsch, O., 1962. Entstehung und Stoffbestand der Salzlagerstätten: Berlin: Springer, 232 p.

Braitsch, O., 1963. Die Entstehung der Schichtung in rhythmisch geschichteten Evaporiten: Geol. Rundschau, v. 52, p. 405-416.

Braitsch, O., 1964. The temperature of evaporite formation, in Problems in palaeoclimatology: London: Interscience, p. 478-490.

Braitsch O., 1971. Salt deposits - their origin and composition: New York: Springer, 297 p.

Briggs, L.I., Jr., 1958. Evaporite facies: Jour. Sed. Petrology, v. 28, p. 46-56.

Briggs, L.I., Jr., 1962. Niagaran-Cayugan sedimentation in the Michigan Basin, in Silurian rocks of the southern Lake Michigan area, Michigan Geol. Soc. Annual Field Conf., Lansing Michigan: Mich. Dept. Conserv., Geol. Survey Div., p. 58-60.

------ 1963. Deposition of evaporites in the Michigan Basin (abs.). in Symposium on Salt: Cleveland, Ohio: Northern Ohio Geol. Soc. p. 55.

------ and Lucas, P.T., 1954. Mechanics of Salina salt deposition in the Michigan Basin (abs.): Geol. Soc. America Bull., v. 65, p. 1233.

Brueren, J.W.R., 1959. The stratigraphy of the Upper Permian Zechstein Formation in the Eastern Netherlands, in I giacimenti gassiferi dell'Europa occidentale: Roma: Accad. Naz. Lincei, v. 1, p. 243-274.

Brunstrom, R.G.W., and Walmsley, P.J., 1969. Permian evaporites in North Sea Basin: Amer. Assoc. Petrol. Geol. Bull., v. 53, p. 870-883.

Butler, G.P., 1970. Trucial Coast - an alternative explanation of origin: Third symposium on salt, Cleveland, Ohio: Northern Ohio Geol. Soc., v. 1, p. 120-152.

------ 1970. Secondary anhydrite from a sabkha, Northwest Gulf of California, Mexico: ibidem, p. 153-156.

Cate, A.S., 1961. Stratigraphic studies of the Silurian rocks of Pennsylvania, Pt. 1: Penn. Geol. Survey, Spec. Bull. 10, 9 p.

------ 1963. Stratigraphy of the Cayugan Series in northwestern Pennsylvania, in Symposium on salt: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 19-26.

------ 1965. Stratigraphic studies of the Silurian rocks of Pennsylvania, Pt. 2: Penn. Geol. Survey, Spec. Bull. 11, 8 p.

Cagle, F.R., Jr., and Cruft, E.F., 1970. Gypsum deposits of the coast of South West Africa: Third symposium on salt, Cleveland, Ohio: Northern Ohio Geol. Soc., p. 156-165.

Carozzi, A.V., 1960. Microscopic sedimentary petrography: New York: Wiley, 485 p.

Chilingar, G.V., et al., 1967. Diagenesis of carbonate rocks, in Diagenesis in sediments - Developments in sedimentology, v. 8: Amsterdam: Elsevier, p. 178-322.

Colton, G.W., 1970. The Appalachian Basin - its depositional sequences and their geologic relationships, in Studies of Appalachian geology - Central and Southern: New York: Interscience, p. 5-47.

Cramer, H.R., 1969. Evaporites - a selected bibliography: Amer. Assoc. Petrol. Geol. Bull., v. 53, p. 982-1011.

Dalrymple, D.W., D.W., 1965. Calcium carbonate deposition associated with blue-green algal mats, Baffin Bay, Texas: Inst. Marine Sci. Pub., v. 10, p. 187-200.

Dellwig, L.F., 1953. Hopper crystals of halite in the Salina of Michigan: Am. Mineralogist, v. 38, p. 730-731.

------1955. Origin of the Salina Salt of Michigan: Jour. Sed. Petr., v. 25, p. 83-110.

------ and Briggs, L.I., Jr., 1952. Textural relationships in the Salina Salt of Michigan (abs.): Geol. Soc. Am. Bull., v. 63, p. 1242.

------ and Evans, R., 1969. Depositional processes in Salina Salt of Michigan, Ohio, and New York: Amer. Assoc. Petrol. Geol. Bull., v. 53, p. 949-956.

Duff, P.McL.D., and Walton, E.K., 1962. Statistical basis for cyclothems - a quantitative study of the sedimentary succession in the East Pennine Coalfield: Sedimentology, v.1, p. 235-255.

------, Hallam, A. and Walton, E.K., 1967. Cyclic sedimentation. Developments in sedimentology 10: Amsterdam: Elsevier, 280 p.

Dunbar, C.O., 1957. Historical geology (2nd ed.): New York: Wiley, 500 p.

Dzens-Litovskiy, A.I., and Vasil'yev, G.V., 1961. Geologic conditions of formation of bottom sediments in Karabogaz-Gol in connection with fluctuations of the Caspian sea level: Izv. Acad. Sci. USSR, Geol. Ser., (Eng. transl.), 1962(3):p79-86.

Eardley, A.J., 1951. Structural geology of North America: New York: Harper, 624 p.

Ehlers, G.M., and Kesling, R.V., 1962. Silurian rocks of Michigan and their correlation, in Silurian rocks of the southern Lake Michigan area: Michigan Basin Ann. Field Conf., Lansing, Mich.: Michigan Dept. Conserv., Geol. Survey Div., p. 1-20.

Elias, G.K., 1963. Habitat of Pennsylvanian algal bioherms, Four Corners area, in Shelf carbonates of the Pradox Basin, a symposium: Four Corners Geol. Soc., 4th Ann. Field Conf., p. 185-203.

Ells, G.D., 1958. Notes on the Devonian-Silurian in the subsurface of southwest Michigan: Michigan Geol. Survey Prog. Rept. 18, 55 p.

------ 1962. Silurian rocks in the subsurface of southern Michigan, in Silurian rocks of the southern Lake Michigan area:Michigan Basin Geol. Soc. Ann. Field Conf., lansing, Michigan: Michigan Dept. Conserv., Geol. Survey Div., p. 39-49.

Evans, G., 1966. The recent sedimentary facies of the Persian Gulf region: Royal Soc. London Philos. Trans., Ser. A, v. 259, p. 291-298.

------ Coastal and nearshore sedimentation - a comparison of clastic and carbonate deposition: Geologists' Assoc. Proc., v. 81, p. 493-508.

Fairbridge, R.W., 1961. Eustatic changes of sea level, in Physics and chemistry of the earth: Oxford: Pergamon, v. 4, p. 99-185.

Fergusson, W.B., and Prather, B.A., 1968. Stratigraphy of the Salina Group in Pennsylvania: Pa. Topog. Geol. Survey, Mineral Resources Rept. M58.

Fettke, C.R., 1955. Preliminary report - rock salt in Pennsylvania: Penn. Geol. Survey Prog. Rept. 145, 4th ser., 1 sheet

Fisher, D.W., 1957. Lithology, paleoecology and paleontology of the Vernon Shale (Late Silurian) in the type area: New York State Mus. Sci. Serv., Bull. 364,31 p.

Fiveg, M.P., 1957. Geologicheskiye usloviya sedimentatsii solenosnykh formatsiy i ikh kaliynykh gorizontov: 20th Internat. Geol. Congr. Proc., sec. 5, v. 1, p. 17-37.

Fuller, J.G.C.M., and Porter, J.W., 1969. Evaporite formations with petroleum reservoirs in Devonian and Mississippian of Alberta, Saskatchewan, and North Dakota: Amer. Assoc. Petrol. Geol. Bull., v. 53, p. 909-926.

Goldsmith, L.H., 1969. Concentration of potash salts in saline basins: Amer. Ass. Petrol. Geol. Bull., v. 53, p. 790-797.

Green, R., 1961. Palaeoclimatic significance of evaporites, in Descriptive palaeoclimatology: New York: Interscience, p.61-68.

Hall, J.F., 1963. Distribution of salt in Ohio, in Symposium on salt: Cleveland, Ohio: Northern Ohio Geol.Soc., p. 27-30.

Herde, W., 1953. Die Riedel-Gruppe im zentralen Teil des nordwestdeutschen Zechsteingebietes: Ph.D. thesis, Göttingen Univ., 127 p.

Hewitt, D.F., 1962. Salt in Ontario: Ontario Dept. Mines Ind. Mineral Rept. 6, 38 p.

Heybroek et al., 1967. Observations on the geology of the North Sea area: 7th World Petroleum Cong. Proc., v. 2, p. 905-916.

Hill, J.V., 1966. Silurian reef carbonates: Ontario Petroleum Inst. 5th Ann. Proc., v. 5, 21 p.

Hite, R.J., 1970. Shelf carbonate sedimentation controlled by salinity in the Paradox Basin, Southeast Utah: Third symposium on salt: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 48-66.

Hofrichter, E., 1960. Zur Stratigraphie, Fazies und Genese der Ronnenberg-Gruppe und Anhydritmittelzone (Zechstein 3) in Nordwest-Deutschland: Ph.D. thesis, Bundesanst. f. Bodenforsch.

Hollingsworth, S.E., 1962. The climatic factor in the geological record: Quart. J. Geol. Soc. London, v. 118, p. 1-21.

Hunter, R.E., Facies of iron sedimentation in the Clinton Group, in Studies in Appalachian geology - Central and Southern:New York: Interscience, p. 101-121.

Ivanov, A.A. and Levitskiy, Yu. F., 1960. Geologiya galogennykh otlozheniy (formatsiy) SSSR: Vsegei, Trudy, new. ser., v. 35.

Jodry, R.L., 1969. Growth and dolomitization of Silurian reefs, St. Clair County, Michigan: Amer. Ass. Petrol. Geol. Bull., v. 53, p. 957-981.

Jacoby, C.H., 1961. Effect of geology on the hydraulic fracturing of salt, in Second symposium on salt: Cleveland, Ohio: Northern Ohio Geol. Soc., v. 2, p. 311-320.

------ 1969. Correlation, faulting, and metamorphism of Michigan and Appalachian Basin salt: Amer. Ass. Petrol. Geol. Bull., v. 53, p. 136-154.

Jung, W., 1959. Das Steinsalzäquivalent des Zechsteins 1 in der Sangerhäuser und Mansfelder Mulde und daraus resultierende Bemerkungen zum Problem der "Jahresringe": Ber. Geol. Ges. DDR, v. 1, p. 313-325.

Jung, W., 1960. Zur Feingliederung des Basalanhydrits (Z 2) und des Hauptanhydrits (Z 3) im SE-Harzvorland: Geologie, v. 9, p. 526-555.

------ 1968. Zechstein; in Abriss der Geologie der DDR, Berlin.

------ 1968. Ueber Gesteinstypen, Faziesdifferenzierungen und zyklisch-rhythmische Sedimentation im deutsch-polnischen Zechstein:XXIII. Intern. Geol. Cong., Sec. 8, Proc., p. 211-225.

Kaufmann, D.W., and Slawson, C.B., 1950. Ripple mark in rock salt of the Salina Formation: Jour. Geol., v. 58, p. 24-29.

Kay, M., and Colbert, E.H., 1965. Stratigraphy and life history: New York: Wiley, 736 p.

Kendall, C.G.St.C., and d'E. Skipwith, P.A., 1968. Recent algal stromatolites of the Khor al Bazam, Abu Dhabi, the southwest Persian Gulf (abs.): Geol. Soc. Amer. Spec. Paper 101, p. 108.

Kent, P.E., 1967. Outline geology of the southern North Sea basin: Yorksh. Geol. Soc. Proc., v. 36, p. 1-22.

Kinsman, D.J.J., 1966. Gypsum and anhydrite of recent age, Trucial Coast, Persian Gulf, in Second symposium on salt, v.1: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 302-326.

------ Modes of formation, sedimentary associations, and diagnostic features of shallow-water and supratidal evaporites: Amer. Ass. Petrol. Geol. Bull., v. 53, p. 830-840.

Klingspor, A.M., 1969. Middle Devonian Muskeg evaporites of Western Canada: Amer. Ass. Petrol. Geol. Bull., v. 53, p. 927-948.

Kosmahl, W., 1969. Zur Stratigraphie, Petrographie, Paläogeographie, Genese und Sedimentation des Gebänderten Anhydrits (Zechstein 2), Grauen Salztones und Hauptanhydrits (Zechstein 3) in Nordwestdeutschland: Hannover: Beih. z. Geol. Jb., v. 71, 129 p.

Kozary, M.T., et al., 1968. Incidence of saline deposits in geologic time, in Saline deposits: Geol. Soc. Amer. Spec. Paper 88, p. 43-57.

Kraus, E.B., 1961. Physical aspects of deduced and actual climatic change: Ann. New York Acad. Sci., v. 95, p. 225-234.

Kreidler, W.L., 1957. Occurrence of Silurian salt in New York State: New York State Mus. and Sci. ServiceBull. 361, 55 p.

------ 1963. Silurian salt of New York State, in Symposium on salt: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 10-18.

Krumbein, W.C., 1951. Occurrence and lithologic associations of evaporites in the United States: Jour. Sed. Petrology, v. 21, p. 63-81.

Kunasz, I.A., 1968. Significance of laminations in the upper Salina Salt of the Michigan basin: M.S. thesis, Penn. State Univ., 76 p.

------ 1970. Significance of laminations in the Upper Silurian evaporite deposit of the Michigan Basin: Third symposium on salt, v. 1: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 67-77.

Landes, K.K., 1945. The Salina and Bass Island rocks in the Michigan basin: U.S. Geol. Survey Oil and Gas Inv. Prelim. Map 40.

------ 1960. The geology of salt deposits, Chapter 4 in Sodium chloride: Amer. Chem. Soc. Monogr. 145: New York: Reinhold, p. 28-69.

------ 1960. Salt deposits of the United States, Chapter 5 in Sodium chloride: Amer. Chem. Soc. Monogr. 145: New York: Reinhold, p. 70-95.

------ 1963. Origin of salt deposits, in Symposium on salt: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 3-9.

Lang, W.B., 1950. Comparison of the cyclic deposits of the Castile and Salado Formations of the Permian of the southwest (abs): Bull. Geol. Soc. Amer., v. 61, p. 1479-1480.

Langbein, R., 1961. Zur Petrographie des Hauptanhydrits (Z3) im Südharz: Chemie der Erde, v. 21, p. 248-264.

Leutze, W.P., 1956. Faunal stratigraphy of Syracuse formation, Onondaga and Madison counties, New York: Amer. Ass. Petrol. Geol. Bull., v. 40, p. 1693-1698.

------ 1960. Stratigraphy and paleontology of the Salina Group in central New York: Ph.D. thesis, Ohio State Univ. (abs.): Dissert. Abs., v. 20, p. 4633-4634.

------ 1964. The Salina Group, in New York State Geol. Assoc. 36th Ann. Mtg. Guidebook: Syracuse, New York: Syracuse Univ. Dept. Geol., p. 57-65.

Lombard, A., 1952. Rythmes sédimentaire et cyclothémes dans lecadre de la sédimentation générale: Congr. Avan. Etudes Stratigr. Géol. Carbonifére, Compte Rendu, 3, Heerlen, v. 2, p. 415-421.

------1956. Géologie sédimentaire - les séries marines: Paris: Masson, 722 p.

Lotze, E., 1938. Steinsalz und Kalisalze, Geologie: Berlin: Borntraeger, 936 p.

------1957. Steinsalz und Kalisalze, 1.Teil, 2. Aufl.: Berlin: Borntraeger, 465 p.

------ 1958. Der englische Zechstein in seiner Beziehung zum deutschen: Geol. Jb., v. 73, p. 135-139.

------ 1964. The distribution of evaporites in space and time, in Problems in palaeoclimatology: London: Interscience, p. 491-507.

Ludlum, J.C., 1959. Rock salt, rhythmic bedding, and salt-crystal impressions in the Upper Silurian limestones of West Virginia: Southeastern Geology, v. 1, p. 22-31.

Lyons, J.B., and Faul, H., 1968. Isotope geochronology of the Northern Appalachians, in Studies of Appalachian geology - Northern and Maritime: New York: Interscience, p. 305-318.

Meckel, L.D., 1970. Paleozoic alluvial deposition in the Central Appalachians- a summary, in Studies in Appalachian geology, Central and Southern: New York: Interscience, p. 49-67.

Melhorn, W.N., 1958. Stratigraphic analysis of Silurian rocks in Michigan basin: Amer. Ass. Petrol. Geol. Bull., v. 42. p. 816-838.

Merriam, D.F., ed., 1964. Symposium on cyclic sedimentation: Kansas Geol. Survey Bull., v. 169, 639 p.

Moore, R.C., Lalicker, C.G., and Fischer, A.G., 1952. Invertebrate fossils: New York: McGraw-Hill, 766 p.

Neev, D., and Emery, K.O., 1967. The Dead Sea - depositional processes and environments of evaporites: Israel Geol. Survey Bull. 41, 147 p.

Parks, W.A., 1908. Silurian stromatoporoids of America: Toronto Univ. Studies, Geol. Ser., v.6, p. 1-52.

Pavlides, L. et al., 1968. Stratigraphic evidence for the Taconic orogeny in the Northern Appalachians, in Studies in Appalachian geology - Northern and Maritime: New York: Interscience, p.61-82.

Pearson, W.J., 1963. Salt deposits of Canada, in Symposium on salt: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 197-239.

Peterson, J.A., and Hite, R.J., 1969. Pennsylvanian evaporite-carbonate cycles and their relation to petroleum occurrence, southern Rocky Mountains: Amer. Ass. Petrol. Geol., Bull., v. 53, p. 884-908.

Phleger, F.B., 1969. A modern evaporite deposit in Mexico: Amer. Ass. Petrol. Geol. Bull., v. 53, p. 824-829.

Poborski, J.W., 1970. The Upper Permian Zechstein in the Eastern Province of Central Europe: Third symposium on salt: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 24-29.

Pray, L.C., and Murray, R.C., ed., 1965. Dolomitization and limestone diagenesis, a symposium: Soc. Econ. Paleontologists Mineralogists, Spec. Publ., v. 13, 88 p.

Richter-Bernburg, G. 1941(2). Zur vergleichenden Stratigraphie des Zechsteins in Mitteldeutschland: Kali 35, p. 193-197

------1950. Zur Frage der absoluten Geschwindigkeit geologischer Vorgänge: Naturwissenschaft, v. 37, p. 1-8.

------ 1955. Stratigraphische Gliederung des deutschen Zechsteins: Z. Deut. Geol. Ges., v. 105, p. 843-854.

------ 1955. Der Zechstein zwischen Harz und Rheinischem Schiefergebirge: Z. Deut. Geol. Ges. 105, p. 876-899.

------ 1957. Isochrone Varven im Anhydrit des Zechsteins 2: Geol. Jb. 74, p. 601-610.

------ 1959. Zur Paläogeographie des Zechsteins, in I giacimenti Gassiferi dell'Europa Occidentale: Roma: Accad. Naz. dei Lincei, v. 1, p. 87-99.

------ 1958. Die Korrelierung isochroner Varven im Anhydrit des Zechsteins 2: Geol. Jb. 75, p. 629-646.

------ 1960. Zeitmessung geologischer Vorgänge nach Varvenkorrelationen im Zechstein: Geol. Rundschau, v. 49, p. 132-148.

------ 1963. Solar cycle and other climatic periods in varvitic evaporites, in Problems in Palaeoclimatology: New York: Interscience, p. 510-519.

Rickard, L.V., 1962. Late Cayugan (Upper Silurian) and Helderbergian (Lower Devonian) stratigraphy in New York: New York State Mus. and Sci. Service Bull. 386, 157 p.

Rickard, L.V., 1966. Upper Silurian Cayugan Series, in Geology of western New York: New York State Geol. Ass. 37th Ann. Mtg.: Buffalo, New York: N.Y. Univ. Dept. Geol. Sci., p. 24-31.

------ 1969. Stratigraphy of the Upper Silurian Salina Group New York, Pennsylvania, Ohio, Ontario: New York State Mus. and Sci. Service, Map and Chart Ser. No. 12, 57 p.

Riley, C.M., and Byrne, J.V., 1961. Genesis of primary structures in anhydrite: Jour. Sed. Petrol., v. 31, p. 553-559.

Rodgers, J., 1968. The eastern edge of the North American continent during the Cambrian and early Ordovician, in Studies of Appalachian geology - Northern and Maritime: New York: Interscience, p. 141-149.

Sanford, B.V., 1965. Salina salt beds southwestern Ontario: Geol. Survey Canada Paper 65-9, 7 p.

Schmalz, R.F., 1966. Environments of marine evaporite deposition: Mineral Industries, v. 35, no. 8, p. 1-7.

------ 1969. Deep-water evaporite deposition - a genetic model: Amer. Ass. Petrol. Geol. Bull., v. 53, p. 798-823.

Scruton, P.C. 1953. Deposition of evaporites: Amer. Ass. Petrol. Geol. Bull., v. 37, p. 2498-2512.

Shapely, H., ed., 1953. Climatic change: Harvard Univ. Press: Cambridge, Mass., 318 p.

Shaver, R.H., 1962. Silurian rocks at the southern edge of the Michigan basin in Indiana, in Silurian rocks of the southern Lake Michigan area, Michigan Basin Geol. Soc. Ann. Field Conf. Lansing, Michigan: Michigan Dept. Conserv., Geol. Survey Div., p. 21-29.

Shaw, A.B., 1964. Time in stratigraphy: New York: McGraw-Hill, 365 p.

Shearman, D.J., 1963. Recent anhydrite, gypsum, dolomite, and halite from the coastal flats of the Arabian shore of the Persian Gulf: Geol. Soc. London, Proc. 1607, p. 63-64.

------ 1966. Origin of marine evaporites by diagenesis: Institution of mining and Metallurgy (Newcastle-upon-Tyne) Trans., Sec. B, v. 75, p. 208-215.

------ 1970. Recent halite rock, Baja California, Mexico: Ibidem, v. 79, p. 155-162.

Simon, P., et al., 1970. Feinstratigraphie, Fazies, und Bromgehalt des Stassfurt-Steinsalzes (Zechstein 2) im Kali- und Steinsalzbergwerk "Asse" (Schacht II) bei Braunschweig: Geol. Jb., v. 88, p. 159-202.

Sloss, L.L., 1953. The significance of evaporites: Jour. Sed. Petrology, v. 23, p. 143-161.

------ 1959. Relationship of primary evaporites to oil accumulation: 5th World Petroleum Cong. Proc., sec. 1, p. 123-138.

------ 1969. Evaporite deposition from layered solutions: Amer. Ass. Petrol. Geol. Bull., v. 53, p. 776-789.

Smith, D.B., 1970. The palaeogeography of the British Zechstein: Third symposium on salt, v. 1: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 20-23.

Soderman, J.W. and Carozzi, A.V., 1963. Petrography of algal bioherms in Brunt Bluff Group (Silurian), Wisconsin: Amer. Ass. Petrol. Geol. Bull., v. 47, p. 1682-1708.

Stehli, F.G., et al., 1963. Evaporite facies in northwestern Ohio, in Symposium on salt: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 31-46.

Stewart, F.H., 1949. The petrology of the evaporites of the Eskdale no.2 boring, east Yorkshire; pt. 1 - the Lower Evaporite Bed: Mineralog. Mag., v. 28, p. 621-675.

------ 1951a. The petrology of the evaporites of the Eskdale no. 2 boring, east Yorkshire; pt. 2 - the Middle Evaporite Bed: Mineralog. Mag., v. 29, p. 445-475.

------ 1951b. The petrology of the evaporites of the Eskdale no. 2 boring, east Yorkshire; pt. 3 - the Upper Evaporite Bed: Mineralog. Mag., v. 29, p. 557-572.

------ 1953. Early gypsum in the Permian evaporites of northeastern England: Geologists' Assoc. (London) Proc. v. 64, p. 33-39.

------ 1954. Permian evaporites and associated rocks in Texas and New Mexico compared with those of northern England: Yorkshire Geol. Soc. Proc., v. 29, p. 185-235.

------ 1956. Replacements involving early carnallite in the potassium-bearing evaporites of Yorkshire: Mineralog. Mag., v. 31, p. 127-135.

Stewart, F.H., 1963a. Marine evaporites, Chapter Y in Data of geochemistry - 6th ed.: U.S. Geol. Survey Prof. Paper 440-Y, p. Y1-Y52.

------ 1963b. The Permian lower evaporites of Fordon in Yorkshire: Yorksh. Geol. Soc. Proc., v. 34, p. 1-44.

------ 1965. The mineralogy of the British Permian evaporites: Mineralog. Mag. (London), v. 34, p. 460-470.

Strakhov, L.N., 1970. Principles of lithogenesis, v. 3: Edinburgh: Oliver & Boyd, 577 p.

Tasch, P., 1963. Fossil content of salt and association evaporites: Symposium on salt: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 96-102.

------ 1970. Biochemical and geochemical aspects of the White Salt Pan - Bonaire, Netherlands Antilles, in Third symposium on salt: Cleveland, Ohio: Northern Ohio Geol. Soc., p. 204-210.

Textoris, D.A., and Carozzi, A.V., 1964. Petrography and evolution of Niagaran (Silurian) reefs, Indiana: Amer. Ass. Petrol. Geol. Bull., v. 48, p. 397-426.

------ and ------ 1966. Petrography of a Cayugan (Silurian) stromatolite mound and associated facies, Ohio: Amer. Ass. Petrol. Geol. Bull., v. 50, p. 1375-1388.

Theokritoff, G., 1968. Cambrian biogeography and biostratigraphy in New England, in Studies of Appalachian geology - Northern and Maritime: New York: Interscience, p. 9-22.

Thompson, R.W., 1970. Tidal-flat sedimentation of the Colorado River delta, northwestern Gulf of California: Geol. Soc. Am. Mem. 107.

Ulteig, J.R., 1964. Upper Niagaran and Cayugan stratigraphy of northeastern Ohio and adjacent areas: Ohio Div. Geol. Survey Rept. Inv. 51, 48 p.

van Straaten, L.M.J.U., 1961 Sedimentation in tidal flat areas: Jour. Alberta Soc. Petroleum Geol., v. 9, p. 203-226.

van Houten, F.B., 1964. Cyclic lacustrine sedimentation, Upper Triassic Lockatong Formation, west-central New Jersey and adjacent Pennsylvania: Am. Jour. Sci., v. 260, p. 561-576.

von Bubnoff, S., 1948. Rhythmen, Zyklen und Zeitrechnung in der Geologie: Geol. Rundschau, v. 35, p. 6-22.

Walther, J., 1912. Das Gesetz der Wüstenbildung in Gegenwart und Vorzeit (2. Aufl.): Leipzig: Quelle & Meyer, 342 p.

Weller, J.M., 1964. Development of the concept and interpretation of cyclic sedimentation: Kansas Geol. Survey Bull., v. 169, p. 607-621.

Willett, H.C., 1961. The pattern of solar climatic relationships: Ann. New York Acad. Sci., v. 95, p. 89-106.

Williams, D., 1961. Sunspot cycle correlations: ibidem, p. 78-88.

Wood, G.W., and Wolfe, M.J., 1969. Sabkha cycles in the Arab/Darb Formation off the Trucial Coast of Arabia: Sedimentology, v. 12, p. 165-191.

Zen, E-an, 1959. Early stages of evaporite deposition (abs.): Geol. Soc. Amer. Bull., v. 70, p. 1704.

------ 1968. Introduction <u>in</u> Studies of Appalachian geology - Northern and Maritime: New York: Interscience, p. 1-5.

------ 1968. Nature of the Ordovician orogeny in the Taconic area: ibidem, p. 129-139.